MINERAL RESOURCES IN THE ARCTIC

The geological surveys active in the Arctic region have compiled information on the most important mineral deposits north of 60°N in a database, on a map and in user-appropriate descriptions for geoscientists in this volume and in a briefer version for general-interest readers. The largest deposits of metals and diamonds on land have been given priority. The briefer version of the description will be published in English, French and Russian.

These products represent the first compilation of information on the most important deposits of the prioritized resource types in the Arctic. The compilation illustrates the importance which the mineral industry has had in the Arctic regions for over a hundred years, but also the priority given to exploration for mineral resources in the region in more recent times. The mineral industry is very important for the northernmost regions of most of the Arctic nations and several deposits north of 70°N (in Canada, Greenland, Norway and Russia) are being mined or developed with a view to mining. Production of certain commodities from mines in the Arctic represents a major part of world production. The results of exploration in both established mining provinces and in new, prospective areas show that there is a considerable potential for new, important discoveries.
MINERAL RESOURCES IN THE ARCTIC
MINERAL RESOURCES
IN THE ARCTIC

Edited by
Rognvald Boyd, Terje Bjerkgård,
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Geological Survey of Norway
PREFACE
Morten Smelror

The Arctic is one of Earth's few remaining frontier areas. This huge area contains some of the largest known provinces of natural resources, including world-class petroleum-bearing basins, metallogenic provinces and mineral deposits, among them several of the world's largest diamond mines. Major new discoveries are still being made in the Arctic, both beyond the regions that are already well known and within provinces where there are already operating mines, where the use of modern exploration methods has revealed "new" ore bodies underneath surface Quaternary deposits or at greater depths in the Earth's crust. Increasing focus on, and development of the Arctic region creates a rapidly growing need for effective assessment of the resource potential of the region. Such an assessment must be based on compilation and evaluation of updated knowledge of the geology of prospective areas and of known deposits.

The Geological Survey of Norway (NGU) held an informal dialogue about possible implementation of a mineral resource project in the Arctic with potential partners in the course of 2011. Having been given a positive response, NGU invited representatives of the geological surveys of the countries of the Arctic region to an inaugural, consultative meeting, held in Toronto in March, 2012 in conjunction with the annual Prospectors and Developers Association of Canada Convention (PDAC). A consensus on the general form of the project was achieved and the first regular meeting of the group of geoscientists who have had a major role in implementation of the project took place in Copenhagen in December 2012. NGU had, prior to that meeting, been granted funding by the Ministry of Foreign Affairs of Norway to support coordination of the project and publication of its products.

Interest in the Arctic will continue to grow in the years ahead of us. Cooperation on compilation of modern geological and geophysical data is a necessary step towards a common understanding of the potential for natural resources in the Arctic regions. This Preface gives me the opportunity to acknowledge the positive support of the leaders of our partner organizations in this project and the outstanding efforts of the geoscientists who have participated in achieving its aims. The financial support of the Ministry of Foreign Affairs of Norway is gratefully acknowledged.

_The fiery skies of the winter, the summer nights' miraculous sun._
_Walk against the wind. Climb mountains._
_Look to the North._
_More often._

Rolf Jacobsen
(from: North in the World, 2002)
## NORWAY

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INTRODUCTION

Geoscientific maps of the Arctic

The first initiative for cooperation on a series of circum-Arctic geological and geophysical maps was undertaken by the Russian Ministry of Natural Resources and Ecology and by the Russian Federal Agency of Mineral Resources (Rosnedra) in 2003 (Petrov and Smelror, 2014). The objective was to produce digital geological and geophysical maps for the Arctic region at a scale of 1:5,000,000. A consortium of national agencies from Canada, Denmark, Finland, Norway, Russia, Sweden and the USA signed an agreement on the cooperation two years later (Petrov & Smelror, 2014). This volume is one of the products from a project, the Circum-Arctic Mineral Resources project, which was among the original aims of the participating organizations and is the most recent to be implemented:

- Geological Map of the Arctic (Harrison et al., 2008)
- Magnetic and Gravity Anomaly Maps of the Arctic (Gaina et al., 2011)
- Mineral Resources in the Arctic (this work)

Compilations of the type provided by the first three Circum-Arctic projects are important steps in updating scientific knowledge, in improving knowledge of the development of the Earth’s Crust and in providing background information, which is relevant in assessment of the potential for mineral resources, not least in relation to energy resources in offshore sedimentary basins.

Why consider the mineral potential of the Arctic?

The High Arctic has attracted attention from explorers at least since the three voyages of Willem Barents in the late 16th C. Barents’ aim was to search for what we now know as the Northeast Passage or, in Russia, as the Northern Sea Route (Figure 1). Barents visited Bear Island, Svalbard and Novaya Zemlya but none of the voyages penetrated the region east of the Kara Sea (Figure 2). The first commercial exploitation in the High Arctic began early in the following century with the establishment of Dutch and English whaling stations on Svalbard. The whaling expeditions discovered coal on Spitsbergen, the main island in the Svalbard archipelago, as early as 1610 and used the coal on their ships (Dallmann, 2015): serious exploration and long-term exploitation of the coal deposits began approximately three hundred years later. Mining of metallic ores has been important in the southern parts of the Nordic region as early as the 9th C (at Falun) and, by the 19th C had become one of the most important industries in northern parts of Norway and Sweden. Mineral exploration and subsequent mining developed throughout the Arctic in the latter part of the 19th C and the first decades of the 20th C, becoming the most important industry in many parts of the Arctic by the latter half of the 20th C. The aim of this work is to give readers an overview of some of the most important mineral deposits on land in the Arctic, not of energy resources such as coal, but of metals and diamonds.

The case for the timeliness of the Circum-Arctic Mineral Resource project rests on many factors:

- Heightened national, regional and international focus on the Arctic including numerous research projects of many kinds.
- National projects on the mineral potential of the Arctic regions of several countries, including Canada, Greenland, the Nordic countries and Russia. These projects involve documentation of mineral potential as part of the basis for assessment of the development potential of the regions covered.
- The continuing discovery of major new deposits in the Arctic, some in known metallogenic
provinces but others in regions not previously recognized as having a major mineral potential.

- Concern relating to access to certain critical mineral resources, some of which are known to occur in the Arctic: assessments of critical raw materials in Europe, the USA and other countries are among the expressions of this concern (European Commission, 2014, US Department of Energy, 2011).

- Improved access due to the more consistent, longer-term opening of shipping lanes such as the North-East Passage (also known as the Northern Sea Route), the North-West Passage and the Arctic Bridge (from Churchill to Murmansk), combined with greater access to ice-classified cargo vessels and ice-breakers.

- Article 7 of the Protocol on Environmental Protection to the Antarctic Treaty (http://www.ats.aq/documents/recatt/att006_e.pdf) which was signed in 1991 bans all mineral resource activities on the Antarctic continent except those related to scientific research. The Arctic Region is thus, on a global scale, one of the few remaining land regions with extensive areas of “prospective” geology in which knowledge of the mineral potential is limited.

Figure 1. Map of the Arctic showing alternative Northwest Passage routes and the Northeast Passage, including the Northern Sea Route along the coast of Siberia (Arctic Marine Shipping Assessment 2009 Report, Arctic Council, April 2009)
Figure 2. The Arctic, according to the four definitions indicated below. (Map courtesy of the The Perry-Castañeda Library Map Collection, University of Texas Libraries, The University of Texas at Austin)
Definition of the Arctic

The following are among several definitions of the Arctic (see Fig. 2):

• The area north of the Arctic Circle, currently (February 2015) defined, using the method developed by Laskar (1986), as 66° 33min 46s north of the Equator. The Arctic Circle is the southernmost latitude in the Northern Hemisphere at which the sun can remain continuously above or below the horizon for 24 hours. Because of changes in the Earth’s axial tilt due to tidal forces, the Arctic Circle at the present moves northwards by about 15 m/year.

• The region in which the average temperature for July is below 10°C.

• The northernmost tree line.

• 60°N. This definition includes:
  Most of Alaska,
  The Yukon, Northwest Territories, Nunavut and the northernmost parts of Québec and Labrador in Canada.
  The whole of Greenland, given a slight infringement of the southern limit in order to reach Cape Farewell at 59° 46min 23s N.
  Iceland
  The Faroe Islands
  The Shetland Islands, except for the southernmost 12 km of Mainland.
  Most of Fennoscandia, approximately the area north of the capitals Oslo, Stockholm and Helsinki.
  Northern Russia, including almost all areas north of the 10°C summer isotherm.

The last of these four definitions is geographically convenient and was adopted in the first of the geoscientific projects implemented in the region.

The land area of the Faroe Islands is not known to contain mineral resources of significance (USGS, 2012) and the metallic mineral deposits known on the northernmost of the Shetland Islands – ophiolite-hosted chromite and platinum group metal deposits are of very limited tonnage (Brough et al., 2015). The deposits on Shetland will not be given further attention in this volume.

The Circum-Arctic Mineral Resource Project

The Fennoscandian Ore-Deposit Database (FODD) project implemented by the Norwegian, Swedish and Finnish geological surveys and authorities in Murmansk and Karelia regions in NW Russia for the area underlain by the Fennoscandian Shield has had great importance for the Circum-Arctic project (http://en.gtk.fi/informationservices/databases/fodd/index.html). The database developed in the FODD project, which is available at the above site, has functioned as a template for the Circum-Arctic project, though the Circum-Arctic project includes, for numerous practical reasons, only the deposits in the three largest size categories of the FODD system.

The method for calculation of the size category of the deposits is described by Eilu et al. (2007). It is based on the calculation of the “in situ” value of each deposit according to ten-year averages of metal prices on the London Metal Exchange (LME) (for the period 1995-2005 in the original version of the database). The values for each deposit are then converted to a tonnage of copper corresponding to the value, so-called copper-equivalent, in order to provide a basis for the comparison of values of widely differing deposits. The FODD database is up-dated every year, most recently in May 2015, using metal prices for the period 2003 – 2012.

This volume is one of the four main products to emerge from the above project. It contains descriptions of the most important metallogenic provinces in the Arctic, the largest individual metal deposits and the most important deposits of diamonds. Metal and diamond deposits were prioritised in the project as being the mineral resources with the highest unit values, as opposed to industrial minerals (in general), construction materials and fertiliser minerals. Information on the individual deposits is entered into a database, which is accessible at the project web site: www.ngu.no/camet. The folder at the back of the volume contains a geological map at a scale of 1:10 million, based on the 1:5 million map (Harrison et al., 2008) on which the largest deposits are plotted.

Inevitably, the authors of the following chapters have faced greatly differing challenges – reflecting the differences in the scale and geological diversity of their countries. The numbers
Figure 3. The Black Angel zinc-lead mine on the W coast of Greenland (photo courtesy of Bjørn Thomassen, GEUS).

Figure 4. Drilling operations at the St. Jonsfjorden gold prospect, Svalbard (photo courtesy of Birger Amundsen, Svalbardposten)
of deposits in the FODD categories Large and Very large is very much larger in certain countries than in others – leading to differences in approach, including greater emphasis on the metallogenic-province level in the former. Independent of such variations the authors have strived to provide the most up-to-date information on the origin and resources of the deposits.

Abbreviated versions of this volume, aimed at general-interest readers, will be published in three languages – English, French and Russian.

REFERENCES


Petrov, O. V., & Smelror, M., 2014: Uniting the Arctic frontiers – International cooperation on circum-Arctic geological and geophysical maps, Polar Record 09, 1-6.


Alaska, the largest State within the United States, and with the majority of Alaska north of latitude 60°, is an important part of the Circum-Arctic region covered by the accompanying metallogenic map. Alaska is a richly endowed region with a long and complex geologic history. The mining history is short by world standards but nevertheless there are a number of world-class deposits in Alaska, of which Red Dog and Pebble are among the largest of their respective types in the world. This chapter provides a brief overview of Alaskan geology, a summary of Alaskan mining history, and a review of the main metallogenic regions of Alaska, with a description of the main deposits within each region. The focus is necessarily on geology and deposits north of latitude 60°N, but for completeness there is a brief summary of geology and major deposits of Alaska to the south of this circum-Arctic boundary. For example, the Pebble deposit is the largest gold resource and one of the largest copper resources in igneous rocks on Earth, but its location at 59° 53' 50'' N in southwestern Alaska is slightly south of the 60° N boundary of the enclosed Circum-Arctic metallogenic map. A brief description of Pebble and the similarly more southerly situated orogenic gold and polymetallic volcanogenic massive sulfide (VMS) deposits of the southeastern Alaska panhandle are included here for completeness.

Alaska consists of an amalgamation of terranes that formed at different times in different places and later came together. Thus, this chapter includes both a simplified introduction to Alaskan tectonic history and, for each major region, a summary of the essential geologic features. We provide an overview of the major ore deposit classes and include more detailed descriptions of individual deposits. Owing to space limitations, we have provided fuller explanations for giant deposits (e.g., Red Dog), but less information for the smaller deposits (e.g., those found in the same Brooks Range region of Alaska). Nonetheless, this chapter is intended to provide an overview of Alaskan geology and the important part that it plays in the metallogeny of the Circum-Arctic region. As was the case with some of the now defined large deposits such as at Fort Knox and Donlin Creek, smaller deposits described here could also eventually be sites of future brownfields exploration and identified economic resources.
Alaska has a geologically complex history that resulted in a collage of terranes or regions having distinct histories, most of which were tectonically assembled from late Paleozoic through Cretaceous (Plafker and Berg, 1994). They now occur as numerous fault-bounded blocks in the northernmost part of the North American Cordillera on the western margin of the Laurentian craton (Colpron et al., 2007). These terranes are comprised of rocks ranging in age from Paleoproterozoic to Recent (enclosed map). The advent of the terrane concept in the 1970s (e.g., Coney et al., 1980) revolutionized the study of geology in Alaska. Initially, many of these terranes were considered exotic (i.e., unrelated to each other and formed in different geologic environments); more recent work has revealed partially common geologic histories and linkages among many of the terranes (e.g., Alexander-Wrangellia-Peninsular composite terrane: Plafker et al., 1994; Beranek et al., 2014; Arctic Alaska, Farewell, and Kilbuck terranes: Bradley et al., 2013). Nonetheless, some of the terranes evolved in widely dispersed locations and were eventually amalgamated to form the present-day State of Alaska.

The oldest rocks in the State are Paleoproterozoic metamorphic rocks. The best documented of these older rocks are located in southwestern Alaska and are part of the Kilbuck terrane. A 2085 Ma zircon from a granitoid in the Kilbuck terrane represents the oldest tightly dated rock in Alaska (e.g., Bradley et al., 2014). These Kilbuck rocks have isotopic signatures that suggest an origin within a continental margin magmatic arc (Miller et al., 1991). Other units that contain detrital and inherited zircons with ages from 2.3 to 2.0 Ga are in east-central Alaska, the southern part of the Brooks Range in northern Alaska, and the Seward Peninsula in northwestern Alaska (Plafker and Berg, 1994), but the rocks themselves are now realized to be much younger (D. Bradley, Bradley Orchards, written commun., 2015).

Geologic evolution of northern and western Alaska: Arctic Alaska and related terranes

The east-west trending Brooks Range of northern Alaska is the northernmost segment of the North American Cordilleran orogen (enclosed map). The regional geology of the Brooks Range has been extensively described by Mull (1982), Moore et al. (1994), and Young (2004). The range is mainly underlain by Neoproterozoic and younger rocks of the Arctic Alaska terrane (Moore et al, 1994), which can be divided in a series of subterranes (Fig. 1) comprising about 25% of Alaska, and is now viewed as a part of the larger Arctic Alaska-Chukotka terrane that includes the Chukotka Peninsula and Wrangel Island of arctic Russia, as well as the North Slope and Brooks Range of Alaska (Fig. 2). The oldest rocks of the Arctic Alaska terrane consist of at least two deformed Proterozoic to Devonian continental margin sequences amalgamated in Late Devonian to Mississippian time (Strauss et al., 2013). Pre-Carboniferous carbonate and siliciclastic rocks in most of the Brooks Range, Seward Peninsula, and adjacent Russia have faunal affinities with Baltica and Siberia (Blodgett et al., 2002; Dumoulin et al., 2002, 2014) and contain detrital zircon populations that suggest proximity with Baltica (Amato et al., 2009; Till et al., 2014); similar-aged rocks in the northeastern Brooks Range have Laurentian origins and may be a fragment of northeastern Laurentia (Strauss et al., 2013). After juxtaposition of these two or more sequences along the Canadian arctic margin, deposition of carbonate platform and fluvial to continental shelf sediments persisted from Late Devonian-Mississippian through Jurassic mainly along the southern side of this presently

Figure 1. Continental margin subterranes of the Neoproterozoic-early Paleozoic Arctic Alaska terrane, part of the Arctic Alaska-Chukotka microcontinent, which were amalgamated along the Canadian Arctic and now partly form the Brooks Range and buried North Slope basement of northern Alaska. Oceanic rocks of the Angayucham terrane were thrust over the rocks of the Arctic Alaska terrane in Early Cretaceous. After Strauss et al. (2013).

Figure 2. Map showing northwestern North America, northeastern Russia, the extent of the Arctic Alaska-Chukotka microcontinent or terrane and locations of select terranes with Neoproterozoic and early Paleozoic rocks. The letter "L" in the northeasternmost part of the Arctic Alaska-Chukotka terrane shows the area where pre-Devonian rocks with Laurentian affinities are exposed (Strauss et al. 2013). Modified after Till et al., 2014.
east-west-trending terrane. Brooks Range orogeny is marked by the Late Jurassic-Early Cretaceous collision of the Koyukuk oceanic arc with the southern subducting margin of the Arctic Alaska terrane (Plafker and Berg, 1994).

The rocks exposed along the southern flank of the Brooks Range are Mississippian and older (Till, 2008) and were those subducted during the early phases of the Late Jurassic or Early Cretaceous orogenesis (Till et al., 1988; Gottschalk, 1990). These blueschist-facies and exhumation-related greenschist-facies rocks of the informally named Schist belt and its overlying carbonate rocks contain volcanogenic massive sulfide and carbonate-hosted copper deposits, respectively (Hitzman et al., 1986). One structural model for the deformed Arctic Alaska terrane rocks north of the Schist belt is that they are a series of seven or more stacked and fault bounded allochthons (Mayfield et al., 1983) resulting from the deformation associated with the arc-continent collision. The structurally lowest of these, the Endicott Mountains allochthon (or subterrane of Fig. 1), is the host for a number of the large shale/mudstone-hosted massive base metal sulfide deposits in the northwestern part of the range.

The arc-continent collision, in addition, emplaced oceanic rocks of the Angayucham terrane structurally above the continental rocks of the Arctic Alaska terrane (Fig. 1). These obducted rocks from a closing ocean basin consist of an upper assemblage of Middle Jurassic peridotite and gabbro and a lower assemblage of Devonian to Jurassic basalt and pelagic sedimentary rocks (Moore et al., 1994). The emplacement of the oceanic rocks, with their contained occurrences of chromite, was coeval with deformation of the Arctic Alaska terrane into a thin-skinned fold-and-thrust belt.

The Seward Peninsula, to the southwest of the Brooks Range, is also underlain by rocks of the Arctic Alaska-Chukotka superterrane or microplate (Fig. 1). The Nome Complex (Fig. 3) of central Seward Peninsula (Till et al., 2011, 2014) consists of a pre-Carboniferous penetratively deformed continental margin sequence that is correlative with protoliths of the southern Brooks Range Schist belt and rocks of the Endicott Mountains allochthon (Till et al., 2014). The Mesozoic deformational history of the Nome Complex parallels that of the Brooks Range Schist belt, as it was also subducted and metamorphosed dur-
The rocks of the Nome Complex host late Early Cretaceous orogenic gold deposits and economically important associated placers. Pre-Devonian non-metamorphosed or slightly metamorphosed carbonate and clastic rocks of the York terrane (Fig. 3) underlie northwestern Seward Peninsula (Dumoulin et al., 2014) and host Late Cretaceous tin granites (Till et al., 2011).

The Farewell terrane in western interior Alaska (Fig. 4), also with Proterozoic basement and a Neoproterozoic to Devonian platform carbonate sequence, is part of the same large early Paleozoic platform that contained the Arctic Alaska terrane somewhere between Laurentia, Siberia, and Baltica. Farewell subsequently rifted apart from the Arctic Alaska-Chukotka microcontinent by Middle Devonian (Bradley et al., 2013). Although lacking any obvious Paleozoic cover rocks, the Paleoproterozoic rocks of the Kilbuck terrane in southwestern Alaska (Fig. 4) were also a part of the microcontinent during its early evolution (Bradley et al., 2013). A vast part of western Alaska, surrounding rocks of the Farewell terrane and south of the Arctic Alaska terrane, is covered by middle to Late Cretaceous terrigenous overlap rocks of the Koyukuk basin and Early Cretaceous andesitic volcanic rocks of the Koyukuk terrane (Patton and others, 1994).

**Accreted terranes of southern Alaska**

South-central Alaska (Fig. 5) mainly comprises the Wrangellia composite terrane and the seaward Chugach terrane that represents a subduction-accretion complex. The Wrangellia composite terrane (or PAW superterrane), underlying parts of present-day southwestern, south-central, and southeastern Alaska, as well as British Columbia and Yukon, Canada, formed during late Paleozoic amalgamation of the Alexander, Wrangellia, and Peninsular oceanic arc terranes; it underlies about 20% of the State (Plafker and Berg, 1994). The formation of the composite terrane occurred in Early Permian between a fringing arc system off western North America and the Uralian orogen that sutured Eurasia and Laurentia (Beranek et al., 2014). By early Mesozoic, Wrangellia was transported far to the south of present-day Alaska along the western margin of Laurentia (Hillhouse and Grommé, 1984; Umhoefer, 2003; Goldfarb et al., 2013).

The oldest rocks of the Chugach terrane were accreted to the seaward side of the Wrangellia composite terrane in early Mesozoic before the latter had collided with the continental margin in the middle Cretaceous. Paleomagnetic results suggest that this initial amalgamation took place south of latitude 25° north (Amato et al., 2013). As the terranes moved to the northeast, they collided with North America, as defined by 101-91 Ma mélange sediments of the Chugach terrane. The majority of the Chugach terrane comprises flysch of the Valdez Group, deposited by
turbidite fans from about 75-55 Ma (Amato et al., 2013; Garver and Davidson, 2015). Seaward of the Chugach terrane, the Prince William terrane represents an early Tertiary continuation of the turbiditic sedimentation seaward of the trench. The Yakutat terrane (included as southern part of Prince William terrane on figure 5 and shown later on figure 33A), the youngest accreted block to the south-central Alaska margin, is a piece of southeastern Alaska and adjacent British Columbia that began to subduct below the accretionary complex at about 30 Ma (Bruhn et al., 2004). The Quaternary volcanoes of the Wrangell Mountains formed along the southeastern side of the Alaska Range in the Quaternary as a consequence of the subduction.

The Border Ranges fault zone is the boundary between the Wrangellia arc terrane and the Chugach accretionary complex. More than 600 km of dextral strike-slip along the fault zone is responsible for much of the deformation of the Chugach complex (Pavlis and Roeske, 2007). Mesozoic arcs (e.g., Talkeetna, Chisana, Chitna, Coast Mountains) formed in the southern part of Wrangelia during subduction of the Chugach oceanic crust. Younger flysch-melt granites were emplaced into rocks of the Chugach and Prince William terranes at 61-48 and 39-29 Ma (Hudson et al., 1979; Plafker et al., 1994). Flysch rocks of the accretionary complex range from low to high metamorphic grades, but most of the rocks of the Valdez Group are metamorphosed to greenschist facies.

**Pericratonic rocks of eastern Alaska**

East-central Alaska (Fig. 6) is defined by pericratonic rocks of the Yukon-Tanana terrane located between the Denali and Tintina strike-slip fault systems, and thus between the seaward Wrangelia and Chugach terranes and Arctic Alaska terrane and related rocks to the north. These pericratonic rocks within the Yukon-Tanana Upland and part of the Alaska Range to the south, which were a part of ancestral North America or Laurentia, have been metamorphosed to greenschist to amphibolite facies. The Neoproterozoic to middle Paleozoic protoliths for the metamorphic rocks were mainly clastic sedimentary rocks, with lesser carbonate and magmatic rocks (Foster et al., 1994). A subduction-related Devonian arc was built upon this attenuating Laurentian continental margin crust, it was rifted in Late Devonian to open the Slide Mountain-Sevemtyme Ocean, and it was accreted back to the
North American margin in the Early Triassic as the ocean basin closed (Nelson et al., 2013). Dusel-Bacon et al. (2013) stressed the composite nature of the Yukon-Tanana terrane/assemblages, with both allochthonous (arc assemblages of figure 6) and parautochthonous (continental margin assemblages of figure 6) components, the latter being the ancestral non-rifted part of the continent margin that hosts the giant Fort Knox and Pogo gold deposits.

Renewed easterly to northeasterly subduction below the Yukon-Tanana terrane is estimated to have occurred from about 220-179 Ma and 115-95 Ma, with at least 400 km of dextral displacement along each of the Tintina and Denali terrane-bounding faults since mid-Cretaceous (Nelson and Colpron, 2007; Allan et al., 2013). The older subduction is associated with emplacement of Early Jurassic arc plutons at depths of >15 km in rocks of the allochthonous assemblage in easternmost Alaska and into adjacent Yukon, Canada (Dusel-Bacon et al., 2013). Subduction beginning at ca. 115 Ma is associated with the extensive northerly translation of rocks outboard of the Denali fault system on the seaward...
side of the Yukon-Tanana terrane. These rocks comprised the amalgamated Paleozoic to Cretaceous Peninsular-Alexander-Wrangellia terrane (Wrangellia composite terrane) and overlapping Middle Jurassic-Early Cretaceous flysch of the remnant ocean basins between Wrangellia and Yukon-Tanana (Umhoefer, 2003; Pavlis and Roeske, 2007). In present-day coordinates, part of the composite terrane and string of remnant basins lies south of Yukon-Tanana in east-central Alaska and defines much of the eastern Alaska Range.

There was widespread intrusion of middle Cretaceous felsic to intermediate, reduced, and peraluminous granitic rocks in the Yukon-Tanana terrane (Hart et al., 2004b). In the Fairbanks area, the plutons occur as shallowly emplaced (<3-5 km), 94-90 Ma isolated domal bodies; in the Goodpaster district, they are more deeply emplaced (5-9 km: Dillworth, 2007) widespread 109-102 Ma batholiths and ca. 93 Ma smaller bodies. Regional metamorphism in Yukon-Tanana is associated with both Early Jurassic contraction and 135-110 Ma extension that exposed much of the parutochthonous assemblage in the westernmost part of the terrane (Hansen and Dusel-Bacon, 1998). To the south, mid-Cretaceous magmatism along the landward side of the Wrangellia composite terrane included emplacement of felsic to intermediate, oxidized, and metaluminous intrusions.

Kuskokwim basin of southwestern Alaska

The 70,000-km² Kuskokwim basin (Fig. 5, 7) underlies much of southwestern Alaska. The basin has been interpreted as a strike-slip basin formed in response to the Late Cretaceous faulting along the Denali-Farewell fault system to the south and Iditarod-Nixon Fork fault system to the north (e.g., Miller and Bundtzen, 1994; Deck er et al., 1994). Most sedimentation took place between 95 to 77 Ma (Miller et al., 2002), when the basin was forming between a series of Middle Jurassic to Early Cretaceous volcanic arc terranes that were approaching the continent from the west and south. To the east of the basin (in present-day configuration), flysch rocks of the slightly older Kahiltna basin (Fig. 5, 7) have an uncertain relationship to the flysch of the Kuskokwim basin (e.g., Graham et al., 2013). Final Kahiltna sedimentation could be as young as 87 Ma (Box et al., 2013). The Kahiltna basin represents the northernmost remnant ocean basin (Fig. 8) closed between Wrangellia and North America (e.g., Hampton et al., 2010). Wrangel lia and perhaps the Kahiltna rocks were south of their present latitude until at least ca. 55 Ma.
Blakely and Umhoefer, 2003). The Kahiltna and Wrangellia rocks have been uplifted to form the present-day western Alaska Range.

The southwestern side of the Kuskokwim basin is bordered by a series of mainly oceanic terranes that were amalgamated with the continent in middle to late Mesozoic (Fig. 7). In addition to the above mentioned old basement rocks of the Kilbuck terrane (Fig. 4), these also include the (1) Late Triassic to Early Cretaceous arc rocks and back-arc volcaniclastics of the Togiak terrane; (2) the Jurassic to Early Cretaceous arc-related volcanic and volcaniclastic assemblage of the Nyac terrane; and (3) the Paleozoic-Mesozoic melange, Permian-Triassic blueschists, and Jurassic mafic and ultramafic rocks of the Goodnews terrane (Decker et al., 1994; Miller et al., 2007). The Cretaceous flysch of the Kuskokwim basin overlaps older rocks of these terranes in many parts of southwestern Alaska.

Most of the clastic rocks in the basins of southwestern Alaska have been metamorphosed to only very low grades. Regional NE-trending fault systems were active during prolonged oblique subduction and host many of the mineral deposits in the region (Bundtzen and Miller, 1997; Graham et al., 2013). Oxidized intermediate to mafic intrusions of the southern Alaska Range plutonic suite of Hart et al. (2004b) were emplaced into both flysch rocks on the seaward side of the Kahiltna basin and rocks of adjacent Wrangellia from about 100-88 Ma. Prior to a second and more widespread magmatic episode beginning at ca. 74 Ma, regional deformation also included development of large E-W-striking fold systems (Miller et al., 2002). Moll-Stalcup (1994) suggested the Campanian and younger magmatism in the Kuskokwim region and adjacent western Alaska Range was part of a Cretaceous-early Tertiary subduction-related continental arc that was as broad as 500 km. In contrast, Bundtzen and Miller (1997) suggested that the young magmatism in the Kuskokwim basin, which was locally alkali-calcic and associated with normal faults, was better classified as back-arc. They also noted a distinct series of similar aged, mainly crustal melt peraluminous granite and alkali-granite sills and dikes that are scattered across the region.
MINING HISTORY OF ALASKA

Except for Alaskan Natives’ utilization of native copper, Alaska’s mining history is relatively short compared to other arctic regions of the world. The earliest attempt by a non-native to mine was in 1848 in south-central Alaska by P.P. Doroshin, a Russian mining engineer sent to southern Alaska from St. Petersburg by the Russian-American Company (Moffit and Stone, 1906). His two-year effort to mine gold was essentially unsuccessful, but later gold rushes opened up much of the State to mining and development. Early prospectors crossed over Chilkoot Pass from coastal Alaska into the Klondike goldfields in Yukon, Canada (Fig. 9), and then eventually into interior Alaska in the 1880s and 1890s.

Perhaps the most spectacular Alaskan gold rush followed announcement in late 1898 of a significant discovery along the beaches of Nome; in 1899 and 1900 (Fig. 10) as many as 20,000 people flocked to this small town along the coast of the Seward Peninsula in northwestern Alaska. The Nome mining district is the second most important placer district in Alaska, having produced more than 155 tonnes (t) Au, essentially all by placer methods and mostly from complex alluvial deposits or buried beach deposits (Metcalf and Tuck, 1942; Bundtzen et al., 1994; Athey et al., 2014). Additional Alaskan placer Au discoveries include the Fairbanks (257 t; Fig. 11), Circle (23 t), Forty Mile (15 t), Hot Springs (14 t), and Tolovana (16 t) districts in interior eastern Alaska and the Iditarod-Flat (45 t) and Innoko (23 t) districts in southwestern Alaska. Follow-up of small concentrations of alluvial gold in the late 1800s and early 1900s led to discovery of the Alaska-Juneau (Fig. 12A, B), Treadwell (Fig. 12C), and Chichagof deposits in southeastern Alaska and the Willow Creek district deposits in south-central Alaska, Alaska’s most significant lode gold producers prior to the 1990s.

Total historic Alaska gold production is likely more than 1400 t, 54 % of which came from placer deposits (Fig. 13). This undoubtedly is a low estimate, as production from small properties often went unreported. In addition, from 1880 through 2013, estimated cumulative Alaskan mining production included about 10,300 t Ag, 1400 t Hg, 5000 t Sb, 3300 t Sn, 2.5 million tonnes (Mt) Pb, 12 Mt Zn, 0.6 Mt Cu, 35,500 t Cr, 600 t U, 0.8 t, and 21 t Pt (Table 1, Athey et al., 2014; http://wr.ardf.usgs.gov).

The late 1890s gold rushes that brought prospectors into Alaska also led to the discovery of the famous high-grade Kennecott-type copper deposits located in the Wrangell Mountains in the easternmost part of south-central Alaska. A railroad was soon built linking the mines with the seaport of Cordova, and production took place between 1911 and 1938 (Fig. 14), which helped lead to the establishment of the Kennecott Copper Corporation. During that period, 4 Mt of ore grading 13 % Cu yielded about 537,000 t Cu, as well as approximately 100 t Ag. The Kennecott district was incorporated into Wrangell-St. Elias National Park in 1980 and the mine workings designated a National Historic Landmark.
Figure 10. Historic placer mining on the rivers and beaches of the Seward Peninsula near Nome led to recovery of more than 155 t Au. A) Locations of historic placer gold workings on the Seward Peninsula, northwestern Alaska, modified from Nelson and Hopkins, 1972. B) View of the beach west of Nome, 1900. Tents and mining equipment are present. Photography by Loment Bros., Nome.
Figure 11. More than 250 t Au were recovered from placer operations in the Fairbanks district, east-central Alaska. A) Hydraulic mining below a 45 m high bank of loess on Cripple Creek, July 31, 1936. B) Fairbanks Gold Mining Company dredge on Fairbanks Creek. Images from Purington (1905).
Base metal deposits were discovered in northwestern Alaska in the late 1960s and exploration took place in the subsequent decades. Iron-oxide staining was first noted along Ferric Creek in the western Brooks Range in 1955 (Koehler and Tikkanen, 1991). Following up on this occurrence, Irv Tailleur, a USGS geologist, sampled stream sediments and rocks along the similar iron-oxide-stained Red Dog Creek (east of Ferric Creek) in 1968 and found >10 % Pb in stream sediments and > 2 % Pb and 1 % Zn in mineralized rock samples (Tailleur et al., 1970). The area was first drilled in 1980 and the second hole intercepted 11.0 m at 48 % Zn and 10 % Pb. Further drilling and ongoing production have established Red Dog as one of the world’s largest clastic-dominated Pb-Zn (SEDEX) deposits, accounting for 4 % of the world’s and 95 % of U.S. zinc reserves (Athey et al., 2014). Other base metal sulfide deposits are recognized in the Red Dog district and
elsewhere in the western Brooks Range, but remain undeveloped.

Additional Alaskan discoveries in the latter half of the 20th century include Quartz Hill in 1974, Greens Creek in 1979, Fort Knox in 1984, Donlin Creek and Pebble (just south of 60°) in 1988, and Pogo in 1994. These deposits are discussed in more detail in later sections. At present Alaska has five active lode mines (Fort Knox, Greens Creek, Kensington, Pogo, and Red Dog) in addition to continuing production from numerous placer gold operations throughout the state. Production data, reported reserves and resources, and total contained metals for the major deposits in Alaska are listed in Tables 2 and 3 for gold and other metals, respectively.

**Figure 13.** Locations of significant placer gold accumulations in Alaska and years of earliest discoveries. Image from Yeend et al. (1998).

**Figure 14.** Operations of the Kennecott copper deposits were active in the Wrangell Mountains from 1911-1938. The old Kennecott mill is now preserved as a National Historic Landmark within Wrangell-St. Elias National Park and Preserve.

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1834 Party of Russian-Americans under Malakoff reports finding gold in the Russian River drainage of the Kenai Peninsula.
1867 Alaska purchased from Russia and officially handed over to the United States in a ceremony at Sitka.
1880 Gold discovered near Juneau, both in the Silver Bow Basin and on Douglas Island.
1886 Gold found in the Fortymile River, the first major gold discovery in the interior of Alaska.
1893 Gold discovered on Birch Creek in an area that later became famous as the Circle Mining District.
1896 George Washington Carmack, Skookum Jim, and Tagish Charlie find rich deposits of gold on a tributary of the Klondike River in the Yukon Territory of Canada, starting the Klondike Gold Rush.
1898 Miners from the Klondike continue down the Yukon to Alaska’s Seward Peninsula and find gold at Nome. Others make finds in other parts of Alaska.
1902 Felix Pedro finds gold on a tributary of the Tanana River at the site what is now the city of Fairbanks.
<table>
<thead>
<tr>
<th>COMMODITY</th>
<th>PRODUCTION</th>
<th>UNITS</th>
<th>VALUE (MILLION $)</th>
<th>YEARS OF PRODUCTION</th>
<th>NOTES</th>
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</thead>
<tbody>
<tr>
<td>Gold</td>
<td>44,904,866 (1,396 t)</td>
<td>Troy ounces</td>
<td>11,496.3</td>
<td>1880 to present</td>
<td>Probably low as reporting is likely incomplete.</td>
</tr>
<tr>
<td>Silver</td>
<td>330,295,964 (10,272 t)</td>
<td>Troy ounces</td>
<td>3,313.2</td>
<td>1880 to present</td>
<td></td>
</tr>
<tr>
<td>Mercury</td>
<td>40,945 (1,415 t)</td>
<td>76 lb flasks</td>
<td>9.9</td>
<td>1920-1980, 1984-1986</td>
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<tr>
<td>Tin</td>
<td>3,312</td>
<td>Tonnes</td>
<td>12.5</td>
<td>1900-1993</td>
<td></td>
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<tr>
<td>Lead</td>
<td>2,488,970</td>
<td>Tonnes</td>
<td>2,993</td>
<td>1880-1980, 1989 to present</td>
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</tr>
<tr>
<td>Zinc</td>
<td>11,980,205</td>
<td>Tonnes</td>
<td>18,541</td>
<td>1940-1949, 1989 to present</td>
<td></td>
</tr>
<tr>
<td>Platinum</td>
<td>673,548 (21 t)</td>
<td>Troy ounces</td>
<td>74.4</td>
<td>1910-1976, 2011</td>
<td>Production data are not available for 1950 to 1976, hence numbers could be quite low. Intermittent low or withheld production 1981-1996</td>
</tr>
<tr>
<td>Uranium, (U₃O₈)</td>
<td>604</td>
<td>Tonnes</td>
<td>n.a.</td>
<td>1957-1991</td>
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</table>

Table 1. Total production – Alaska 1880 to 2013 (Athey et al., 2014), Uranium from ARDF data (http://wr.ardf.usgs.gov).

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<tr>
<th>DEPOSIT</th>
<th>LATITUDE (N)</th>
<th>LONGITUDE (W)</th>
<th>DISTRICT</th>
<th>DEPOSIT TYPE</th>
<th>YEARS OF PRODUCTION</th>
<th>AU PRODUCTION (TONNES)</th>
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<tr>
<td>Alaska-Juneau</td>
<td>58.308</td>
<td>134.342</td>
<td>Juneau Gold Belt</td>
<td>Orogenic</td>
<td>1883-1944</td>
<td>95.5</td>
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<tr>
<td>Apollo</td>
<td>55.191</td>
<td>160.563</td>
<td>Shumagin Islands</td>
<td>Epithermal</td>
<td>1892-1904, 1908-1913</td>
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<td>Chichagof</td>
<td>57.663</td>
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<td>Chichagof</td>
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<td>1906-1942</td>
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<td>Cleary Hill</td>
<td>65.067</td>
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<td>Fairbanks</td>
<td>Orogenic</td>
<td>1910-1915, 1929-1932</td>
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<tr>
<td>Donlin Creek</td>
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<td>Kuskokwim</td>
<td>Orogenic(?)</td>
<td>undeveloped</td>
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<td>Fort Knox</td>
<td>64.992</td>
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<td>Treadwell</td>
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Note: Data from USGS Alaska Resource Data Files and other references cited

Table 2. Past production and reserves/resources for significant lode gold deposits in Alaska.
## Commodity Production

### Units and Value

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<th>COMMODITY</th>
<th>PRODUCTION UNITS</th>
<th>VALUE (MILLION $)</th>
<th>YEARS OF PRODUCTION</th>
<th>NOTES</th>
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<tr>
<td>Gold</td>
<td>44,904,866</td>
<td>(1,396 t)</td>
<td>1880 to present</td>
<td>Probably low as reporting is likely incomplete.</td>
</tr>
<tr>
<td>Silver</td>
<td>330,295,964</td>
<td>(10,272 t)</td>
<td>1880 to present</td>
<td></td>
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<tr>
<td>Mercury</td>
<td>40,945</td>
<td>(1,415 t)</td>
<td>1920-1980, 1984-1986</td>
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<td>Tin</td>
<td>3,312 Tonnes</td>
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<td>Lead</td>
<td>2,488,970 Tonnes</td>
<td>2,993</td>
<td>1880-1980, 1989 to present</td>
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<td>Zinc</td>
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<td>18,541</td>
<td>1940-1949, 1989 to present</td>
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<tr>
<td>Platinum</td>
<td>673,548 Tonnes</td>
<td>74.4</td>
<td>1910-1976, 2011</td>
<td>Production data are not available for 1950 to 1976, hence numbers could be quite low. Intermitent low or withheld production 1981-1996</td>
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<tr>
<td>Uranium</td>
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### Deposit Data

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<th>DEPOSIT TYPE</th>
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<th>AU PRODUCTION (TONNES)</th>
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<th>OTHER METALS PROD. (TONNES)</th>
<th>GOLD RESERVES (TONNES)</th>
<th>GOLD RESOURCE (TONNES)</th>
<th>GROSS GOLD (GRAMS/TONNE)</th>
<th>REFERENCES</th>
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<td>Alaska-Juneau</td>
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<td>Juneau Gold Belt</td>
<td>Orogenic</td>
<td>1883-1944</td>
<td>95.5</td>
<td>1.6</td>
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<td>24</td>
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<td>158.209</td>
<td>Kuskokwim</td>
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<td>147.361</td>
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<td>1996-present</td>
<td>181</td>
<td>0.9</td>
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<td>Independence</td>
<td>61.792</td>
<td>149.294</td>
<td>Willow Creek</td>
<td>Orogenic</td>
<td>1909-1942, 1949-1951</td>
<td>5.3</td>
<td>approx 35</td>
<td>5.5 t Ag</td>
<td>minor</td>
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<tr>
<td>Kensington</td>
<td>58.864</td>
<td>135.082</td>
<td>Juneau Gold Belt</td>
<td>Orogenic</td>
<td>1897-1900, 2010-present</td>
<td>12.6</td>
<td>5.8</td>
<td>18</td>
<td>27</td>
<td>5.8</td>
<td><a href="http://www.coeur.com">www.coeur.com</a></td>
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<tr>
<td>Lucky Shot/War Baby</td>
<td>61.779</td>
<td>149.408</td>
<td>Willow Creek</td>
<td>Orogenic</td>
<td>1922-1942</td>
<td>7.1</td>
<td>possibly Te</td>
<td>minor Cu</td>
<td>minor</td>
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<tr>
<td>Money Knob</td>
<td>65.509</td>
<td>148.534</td>
<td>Livengood/Tolovana</td>
<td>Intrusion-related</td>
<td>undeveloped</td>
<td>570</td>
<td>0.58</td>
<td>570</td>
<td>0.58</td>
<td><a href="http://www.ithmines.com">www.ithmines.com</a></td>
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<tr>
<td>Pogo</td>
<td>64.453</td>
<td>144.914</td>
<td>Goodpaster</td>
<td>Orogenic</td>
<td>2006-present</td>
<td>approx 90</td>
<td>14</td>
<td>51</td>
<td>77</td>
<td>11.5</td>
<td>Freeman (2015)</td>
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<td>Treadwell</td>
<td>58.269</td>
<td>134.378</td>
<td>Juneau Gold Belt</td>
<td>Orogenic</td>
<td>1882-1923</td>
<td>91.6</td>
<td>3.6</td>
<td>5.1 t Ag</td>
<td>significant</td>
<td>Redman et al. (1991)</td>
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<td>DEPOSIT</td>
<td>LATITUDE (N)</td>
<td>LONGITUDE (W)</td>
<td>DEPOSIT TYPE</td>
<td>YEARS MINING</td>
<td>METAL GRADES</td>
<td>RESERVE (CONTAINED METAL) GRADE</td>
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<tr>
<td>Anarraaq</td>
<td>68.155</td>
<td>163.033</td>
<td>Clastic-dominated Pb-Zn</td>
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<tr>
<td>Arctic</td>
<td>67.174</td>
<td>156.3875</td>
<td>VMS</td>
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</tr>
<tr>
<td>Beatson</td>
<td>60.0493</td>
<td>147.8995</td>
<td>VMS</td>
<td>1908-1930</td>
<td>0.083 Mt Cu, 0.13 Mt of Pb</td>
<td>1.65% Cu, 8.7 g/t Ag</td>
<td></td>
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</tr>
<tr>
<td>Bond Creek</td>
<td>42.198</td>
<td>142.7027</td>
<td>Porphyry</td>
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<tr>
<td>Bornite (Ruby Creek)</td>
<td>67.0624</td>
<td>156.948</td>
<td>Carb. Copper</td>
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<tr>
<td>Brady Glacier</td>
<td>58.5533</td>
<td>136.9275</td>
<td>Magmatic Ni-Cu</td>
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<tr>
<td>Death Valley (Boulder Ck)</td>
<td>65.0507</td>
<td>162.2467</td>
<td>Sed. Uranium</td>
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<tr>
<td>Greens Creek</td>
<td>58.079</td>
<td>134.6312</td>
<td>VMS</td>
<td>1989-1993, 1996-present</td>
<td>1.36 Mt Zn, 0.43 Mt Pb, 0.89 Mt of Cu, 0.54 Mt of Pb</td>
<td>8.7% Zn, 3.3% Pb, 2.8 g/t Au</td>
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<tr>
<td>Kemuk (Humble)</td>
<td>59.7203</td>
<td>157.67</td>
<td>Urals-Alaska type complex</td>
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<tr>
<td>Kennebecott</td>
<td>61.522</td>
<td>142.821</td>
<td>Kennebecott-type copper</td>
<td>mainly 1911-1938</td>
<td>0.56 Mt Cu, approx 100 t Ag</td>
<td>13% Cu</td>
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<tr>
<td>Klukwan</td>
<td>59.42</td>
<td>135.88</td>
<td>Urals-Alaska type complex</td>
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</tr>
<tr>
<td>Lakeview/Longview</td>
<td>68.6066</td>
<td>157.4783</td>
<td>Sed. Barite</td>
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<td></td>
</tr>
<tr>
<td>Lost River</td>
<td>65.474</td>
<td>167.156</td>
<td>Skarn</td>
<td>1952-1956</td>
<td>314 t Sn</td>
<td>1.42% Sn</td>
<td></td>
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</tr>
<tr>
<td>Nixon Fork</td>
<td>63.2381</td>
<td>154.7759</td>
<td>Skarn</td>
<td>1918-1964, 1995-1999, 2007, 2011-2013</td>
<td>6.1 t Au, 1200 t Cu, approx 100 t Ag</td>
<td>0.9% Cu</td>
<td></td>
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<tr>
<td>Orange Hill</td>
<td>42.204</td>
<td>142.8449</td>
<td>Porphyry</td>
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</tr>
<tr>
<td>Pebble</td>
<td>59.8991</td>
<td>155.2807</td>
<td>Porphyry</td>
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<td></td>
</tr>
<tr>
<td>Quartz Hill</td>
<td>55.403</td>
<td>130.483</td>
<td>Porphyry</td>
<td></td>
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<td></td>
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</tr>
<tr>
<td>Red Dog</td>
<td>68.0704</td>
<td>162.8379</td>
<td>Clastic-dominated Pb-Zn</td>
<td>1989-present</td>
<td>11.3 Mt Zn, 2.0 Mt Pb, 0.18 Pt, 0.27 Mo</td>
<td>20% Zn, 6% Pb, 38-43% Cr, 0.3 g/t Ag</td>
<td></td>
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<tr>
<td>Red Devil</td>
<td>61.75947</td>
<td>157.31375</td>
<td>Epizonal Hg-Sb</td>
<td>1933-1946, 1953-1963, 1969-1971</td>
<td>125 t Ag, 25 g/t Ag, 27 t Ag</td>
<td>6.5% Zn, 4% Pb, 0.5% Cu</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Red Mountain (Dry Creek)</td>
<td>63.92</td>
<td>147.38</td>
<td>VMS</td>
<td></td>
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<td></td>
<td></td>
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<tr>
<td>Red Mountain</td>
<td>59.35</td>
<td>151.49</td>
<td>Magmatic chromite</td>
<td>1920-1942, 1958</td>
<td>approx 14,000 t Cr, 0.3 g/t Ag</td>
<td>38-43% Cr, 0.3 g/t Ag</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Whistler</td>
<td>61.9638</td>
<td>152.6799</td>
<td>Porphyry</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>WTF</td>
<td>63.92</td>
<td>147.38</td>
<td>VMS</td>
<td></td>
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Note: Presently active mine production data estimated through end of 2014; resources include inferred, except for Red Dog and Greens Creeks where they are measured/indicated; data from USGS Alaska Resource Data Files and other references cited.

Table 3. Resources in base metal deposits and occurrences in Alaska.
<table>
<thead>
<tr>
<th>RESOURCE (CONTAINED METAL)</th>
<th>METAL GRADES</th>
<th>GOLD T</th>
<th>SILVER T</th>
<th>ZINC MT</th>
<th>LEAD MT</th>
<th>COPPER MT</th>
<th>CU EQUIV MT</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.7 Mt Zn, 0.82 Mt Pb, 1125 t Ag, 1000 Mt barite</td>
<td>15.8% Zn, 4.8% Pb, 65 g/t Ag</td>
<td>1125</td>
<td>2.7</td>
<td>0.82</td>
<td></td>
<td></td>
<td></td>
<td>Athey et al. (2014)</td>
</tr>
<tr>
<td>0.89 Mt Cu, 1.19 Mt Zn, 0.2 Mt Pb, 177 t Au, 1285 t Ag</td>
<td>3.3% Cu, 4.5% Zn, 0.8% Pb, 0.7 g/t Au, 53 g/t Ag</td>
<td>177</td>
<td>1285</td>
<td>1.19</td>
<td>0.2</td>
<td>0.89</td>
<td></td>
<td>Wilkins et al. (2013)</td>
</tr>
<tr>
<td>some ore reported remaining</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>41.5</td>
<td>0.083</td>
</tr>
<tr>
<td>1.5 Mt Cu, minor Ag, Au, Mo</td>
<td>0.30% Cu</td>
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<td></td>
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<tr>
<td>2.74 Mt Cu</td>
<td>1.6% Cu</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2.74</td>
</tr>
<tr>
<td>0.46 Mt Ni, 0.27 Mt Cu, 71 PGEs</td>
<td>0.5% Ni, 0.3% Cu, 0.18 ppm PGE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.27</td>
</tr>
<tr>
<td>4531 U₂O₆</td>
<td>0.27% U₂O₆</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>0.051 Mt Zn, 0.023 Mt Pb, 266 t Ag, 2 t Au</td>
<td>7.3% Zn, 3.3% Pb, 381 g/t Ag, 2.8 g/t Au</td>
<td>64</td>
<td>968</td>
<td>2.03</td>
<td>0.68</td>
<td></td>
<td></td>
<td>Athey et al. (2014)</td>
</tr>
<tr>
<td>384 Mt Fe, also Ti, P</td>
<td>15-17% Fe</td>
<td></td>
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</tr>
<tr>
<td>535 Mt Fe, also Pt, Pd</td>
<td>16.8% Fe</td>
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<tr>
<td>33.2 Mt barite</td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>54,600 t Sn</td>
<td>0.26% Sn</td>
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<tr>
<td>4.8 t Au</td>
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<td></td>
<td></td>
<td>10.9</td>
</tr>
<tr>
<td>1.36 Mt Cu, 45 t Au, 90 t Ag, minor Mo</td>
<td>0.35% Cu</td>
<td>45</td>
<td>90</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.36</td>
</tr>
<tr>
<td>37 Mt Cu, 2.5 Mt Mo, 3033 t Au, 14,572 t Ag</td>
<td>0.24-0.42% Cu, 0.26-0.35 g/t Au, 1.19-1.66 g/t Ag, 215-250 ppm Mo</td>
<td>3033</td>
<td>14572</td>
<td></td>
<td></td>
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<td>37 <a href="http://www.northerndynastyminerals.com">www.northerndynastyminerals.com</a></td>
</tr>
<tr>
<td>2.03 Mt Mo</td>
<td>0.127% Mo</td>
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<tr>
<td>1.9 Mt Zn, 0.57 Mt Pb, 1038 t Ag</td>
<td>25.7% Zn, 6.9% Pb, 125 g/t Ag</td>
<td>7738</td>
<td>21.1</td>
<td>5</td>
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<td>Athey et al. (2014)</td>
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<tr>
<td>0.13 Mt Zn, 0.055 Mt Pb, 273 t Ag, 1.5 t Au</td>
<td>4.4% Zn, 2.5% Pb, 94 g/t Ag, 0.5 g/t Au</td>
<td>1.5</td>
<td>273</td>
<td>0.13</td>
<td>0.055</td>
<td></td>
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<td>Dusel-Bacon et al. (2010)</td>
</tr>
<tr>
<td>1.36 Mt Cr₂O₃</td>
<td>mainly 5-6% Cr₂O₃</td>
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</tr>
<tr>
<td>89 t Au, 375 t Ag, 0.35 Mt Cu</td>
<td>0.44 g/t Au, 1.8 g/t Ag, 0.16% Cu</td>
<td>89</td>
<td>375</td>
<td>0.35</td>
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<td><a href="http://www.brazilresources.com">www.brazilresources.com</a></td>
</tr>
<tr>
<td>0.17 Mt Zn, 0.07 Mt Pb, 501 t Ag, 2.5 t Au</td>
<td>6% Zn, 2.5% Pb, 179 g/t Ag, 0.9 g/t Au</td>
<td>2.5</td>
<td>501</td>
<td>0.17</td>
<td>0.07</td>
<td></td>
<td></td>
<td>Dusel-Bacon et al. (2007)</td>
</tr>
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</table>

**TOTAL METAL (COMBINED PRODUCTION, RESERVES, AND RESOURCES)**
Northern Alaska

Clastic-dominated Pb-Zn (SEDEX) deposits
Clastic-dominated Pb-Zn deposits are those traditionally referred to as sedimentary-exhalative (SEDEX), many of which however lack definitive evidence of exhalative processes (Leach et al., 2010). The Pb-Zn ores are mainly hosted in the structurally lowest Endicott Mountains allochthon or subterrane in the western Brooks Range (Fig. 15A). They mostly occur in relatively non-metamorphosed, fine-grained Mississippian clastic rocks, turbiditic carbonate rocks, and chert at the top of the Kuna Formation within the Lisburne Group (Fig. 15B). The clastic rocks include black siltstone, and siliceous and carbonateous mudstone and shale. The upper part of the Kuna Formation that hosts all sulfide deposits in the Red Dog district has a total thickness of 30-240 m (Dumoulin et al., 2004). Deposition occurred in a late Early to Late Mississippian anoxic to euxinic basin isolated from the open ocean (Johnson et al., 2015) with limited siliciclastic input and significant amounts of organic carbon; carbonate turbidites derived from adjacent carbonate platforms are locally present in the Kuna Formation (Dumoulin et al., 2004).

Based on contained Zn+Pb, Red Dog (Very Large) is one of the world’s largest clastic-dominated Pb-Zn type deposits. The mine at Red Dog has recovered ore from two orebodies or deposits: the Main deposit (mined out in 2012) and the Aqqaluk deposit (currently being mined). The Qanaiyaq (or Hilltop) and Paalaaq deposits are potential sources of near-term higher-grade ore to supplement the reserves currently being mined from the adjacent Aqqaluk pit (Fig. 16A, B) (Athey et al., 2014). Collectively, the four orebodies are referred to as the Red Dog deposits of Kelley and Jennings (2004). The four deposits at Red Dog have a cumulative reserve and resource of 140.6 Mt of 16.6 % Zn and 4.6 % Pb. Also within the broader Red Dog district are important unmined resources of Zn+Pb at Su-Lik (Very Large) and Anarraaq (Very Large). A recent estimate indicates a combined pre-mining estimate for all district deposits of 171 Mt containing 15.7 % Zn, 4.5 % Pb, and 82.6 g/t Ag (Blevings et al., 2013). Numerous barite bodies, some associated with the Zn-Pb deposit, are scattered throughout the district and include an estimated 1000 Mt of barite at Anarraaq.

Ore minerals in the Zn-Pb-Ag deposits of the Red Dog district include sphalerite, galena, pyrite, and marcasite (Kelley et al., 2004b). Copper-bearing sulfide phases are rare. Mineralization styles for the base metal sulfides include vein (Fig. 16C), massive (Fig. 16D), breccia, and disseminated. The Anarraaq deposit northwest of Red Dog formed by replacement of carbonate turbidites that are present as layers within the mudstone sequence. The Red Dog deposit ores are very coarse-grained and may be brecciated, whereas deposits such as Anarraaq and Lik-Su are predominantly characterized by extremely fine-grained sulfide layers. Due to post-mineralization deformation during the Brookian orogeny, the Red Dog deposits are structurally separated in a series of thrust slices of siliceous shale and chert. A paragenesis for ore formation is detailed as follows: (1) dominantly barite, with volumetrically minor brown sphalerite and minor galena and pyrite deposition in unconsolidated mud below the ocean floor; (2) yellow-brown sphalerite, galena, and pyrite deposition and recrystallization and coarsening of early barite; (3) red-brown sphalerite, with galena, pyrite, and marcasite, deposited in veins and replacing older barite; and (4) deposition of post-ore sulfides and formation of tan sphalerite breccias (Kelley et al., 2004b). The bulk of the Red Dog ore is the massive to semi-massive yellow-brown sphalerite of the 2nd stage, although 20 % of the zinc ore is in veins as red-brown sphalerite. Kel-

1 The sizes of the deposits described in this chapter are grouped according to the criteria described by Eilu et al. (2007).
ley et al. (2004b) described a Zn:Pb metal zoning in the Red Dog deposits of 4.5 at the base to 3.0 at the top. Although pyrite is not particularly abundant, Slack et al. (2004) indicated that it contained exceptionally high amounts of TI, locally as great as 1.2%. Graham et al. (2009) suggested that TI may serve as the best geochemical pathfinder in exploration for clastic-dominated Pb-Zn deposits in the western Brooks Range.

Barite and quartz are the main gangue phases in the Red Dog ore. Minor calcite and apatite also have been observed. Quartz is atypically abundant in the Red Dog deposits relative to most clastic-dominated Pb-Zn deposits. Leach et al. (2004) argued that the widespread silicification occurred during the Brookian orogeny, ca. 200 m.y after ore deposition, although Slack et al. (2004) argued some of the quartz was pre- and syn-ore.

Fluid inclusion studies by Leach et al. (2004) showed ore fluid salinities as high as 30 wt. % NaCl equiv. in supratidal evaporative brines,
which were generated at a paleolatitude much lower than that of present day Alaska (Lewchuk et al., 2004). The fluid inclusion data also showed ore deposition between 100 and 200°C and at no more than 150 bars, before burial to depths of as much as 12 km during Mesozoic orogeny. Leach et al. (2004) pointed out that these data differ from all previous fluid inclusion data from Red Dog, which indicated hotter, deeper, and less saline fluids, because previous studies did not recognize that much of the studied quartz was not part of the Carboniferous ore event.

Pyrite from the main stage of mineralization at Red Dog was dated by a 10-point Re-Os isochron of 338.3±5.8 Ma (Morelli et al., 2004). This is roughly the age of deposition of the Osagean to early Chesterian (ca. 347-330 Ma) ore host rocks (Dumoulin et al., 2004). Thus, although ore textures clearly show replacement textures, such replacement occurred soon after sedimentation. Rhenium-Os isotopic measurements on sphalerite were not successful probably because of disturbances of the system by later hydrothermal events and/or the Brookian deformation. Such hydrothermal overprinting is further suggested by Jurassic-Cretaceous 40Ar/39Ar ages of hydrothermal ore-related white micas from the Red Dog and Anarraaq deposits (Rombach and Layer, 2004).

The barite bodies mentioned above may or may not be spatially or temporally associated with the massive sulfides. In some places within the district, barite and sulfide orebodies are spatially superimposed (e.g., Red Dog deposits) and at others they are spatially distinct (e.g., Anarraaq deposit contains barren barite separated from the

underlying sulfide orebody by ~70 m of unmineralized black shale). In all studied examples, barite formed prior to sulfide mineralization (Kelley and Jennings, 2004). All the stratiform barite deposits recognized in the Red Dog district are present along the contact between the fine-grained clastic rocks of the Carboniferous Kuna Formation and the mid-shelf or deeper shale and chert of the Late Carboniferous Siksikpuk Formation (Kelley et al., 2004a). The barite bodies, mainly hosted by the limestone and chert along the contact, are as thick as 145 m. Narrow veins and lenses of barite are also present up-section in sedimentary rocks as young as Triassic. The stratiform barite in the Red Dog district formed during a brief period of ventilation and oxidation of the deep Kuna basin late in its overall history but prior to sulfide formation (Johnson et al., 2015). In contrast, Reynolds et al. (2015) argued that the ore-hosting sediments were deposited in a shelf setting in which redox conditions were affected by a fluctuating oxygen minimum zone and that trace fossils may have played an important role in controlling the flow of ore-forming fluids by increasing host sediment permeability.

The Pb-Zn deposits of the Red Dog district formed in large part by replacement of sub-seafloor strata (Kelley et al., 2004b; Leach et al., 2004, 2005), and not exhalation on the seafloor (e.g., Moore et al., 1986). Extensional tectonism likely initiated fluid flow beneath the isolated basin floor. Metals were scavenged by brines from Devonian to Mississippian underlying clastic rocks of the Endicott Group or deeper fractured basement rocks and the metalliferous fluids moved upward along the extension-related faults (Leach et al., 2004). Zinc-Pb-Ag ores were subsequently deposited in permeable parts of the stratigraphy where the infiltrating fluids were H_2S-rich (Johnson et al., 2015). The H_2S was supplied by reduction of pore water sulfate or dissolution of pre-existing barite (Kelley et al., 2004b). Sulfide deposition was mainly via fracture filling and replacement within the calcareous or carbonaceous rocks and the barite.

Similar, but smaller Zn-Pb-Ag deposits are exposed to the east of the Red Dog district for slightly more than 300 km (e.g., Werdon et al., 2004). The most significant of these is Drenchwater, located 140 km east of the Red Dog district and also hosted by rocks of the Kuna Formation (Fig. 15). Unlike the Red Dog district, igneous rocks are present in the area of the Drenchwater deposit and include trachyte, andesite and basalt, which led workers to originally suggest the deposit was a VMS (e.g., Nokleberg and Winkler, 1982). Barite is absent at Drenchwater (Werdon, 1996). Disseminated mineralization is continuous for 3 km along strike and more local semi-massive mineralization is mottled and brecciated (Schmidt, 1997). Ore minerals are light to brown sphalerite, galena, marcasite, and pyrite in a siliceous matrix.

In addition to the occurrences in mudstone and shale, banded sulfide-rich Zn-Pb-Ag veins and breccias are hosted by more coarse-grained clastic lithologies (e.g., Kelley et al., 1997), which are lower in the stratigraphy than rocks of the Kuna Formation. In such vein-breccias, higher grades of copper are present. Also throughout this belt to the east of Red Dog, more distal mounds of barite (Fig. 17) are scattered across the southwestern Brooks Range, such as at Lakeview (Fig. 15, Table 3).

Volcanogenic massive sulfide deposits
The Ambler district (Fig. 18), hosted in the Schist belt, contains a series of VMS deposits that continue for more than 100 km along strike, parallel to the southern margin of the range (Hitzman et al., 1986). The deposits are associated with Middle to Late Devonian bimodal volcanic rocks, particularly schistose metarhyolites, that were likely rift-related and hence are older than the Pb-Zn deposits discussed above. Hitzman et al. (1986) argued that the presence of shallow-wa-
ter fossils associated with carbonate rocks and volcanic units, most likely formed subaerially, were consistent with ore formation in shallow environments. The VMS deposits are hosted by greenschist facies rocks of the Ambler Sequence of Hitzman et al. (1982), which is dominantly a calcareous and volcaniclastic sedimentary unit that contains abundant rhyolite and basalt flows and tuff. The lowest of three tuff units in the sequence, the 80-m-thick Arctic tuff (Hitzman et al., 1986), hosts the majority of the VMS mineralization.

Arctic (Very Large) is the most significant of these Schist belt deposits (Schmidt, 1986), although none of the deposits have any production. The Arctic deposit (Fig. 19) is being explored by NovaCopper Inc. and has an indicated and inferred resource estimate of about 27 Mt of 3.3 % Cu, 4.5 % Zn, 0.8 % Pb, and 53 g/t Ag (Wilkins et al., 2013). The deposit is located between two metamorphosed rhyolite flow dome complexes (Schmidt, 1986). The orebodies consist of a number of semi-massive sulfide lenses that are up to 14-18 m thick and are situated along the limbs of an anticline within the schist belt (Newberry et al., 1997b; Wilkins et al., 2013). The main mineralized area covers about 1.3 x 1 km and extends to a depth of at least 250 m. Most of the ore appears as replacement style mineralization; stringers and stockworks are lacking (Wilkins et al., 2013). Uranium-Pb dating of syn-mineralization metaryolites indicates ore deposition took place between 382 and 373 Ma (Ratterman et al., 2006). Geochemical data for the dated rocks and associated metabasalts suggest that the ore-forming event took place in a back-arc setting (Ratterman et al., 2006).

The ore minerals include coarse-grained chalcopyrite, sphalerite, galena, tetrahedrite-tennantite, arsenopyrite, and minor pyrite and pyrrhotite. High silver grades are associated with the tetrahedrite-tennantite. Less common ore-related phases include breithauptite, bornite, carrollite, covellite, cubanite, digenite, electrum, enargite, glaucodot, and stromeyerite (Schmidt, 1988). Schmidt (1983) described footwall Mg-rich chloritic alteration, a hangingwall pyrite-quartz assemblage, and an assemblage of phlogopite, barium-rich phengite, talc, calcite, dolomite, quartz, and barite interlayered with the sulfide horizons. Mineralization and alteration studies, as described by Schmidt (1986), in-
dicate that metals are zoned laterally from Cu-, to Zn-, and finally to Pb-Ag-rich assemblages at increasing distance from the central chloritic alteration zone. Barite gangue is specifically associated with the zinc mineralization. There is, however, no obvious vertical zoning.

Other smaller, but similar VMS occurrences within the Schist belt include Dead Creek (Shungnak) Sunshine, Horse Creek (Cliff), Sun, Tom-Tom, and Smucker (Hitzman et al., 1986; Schmidt, 1988; Newberry et al., 1997b). Most of these are Cu-Zn occurrences, but Horse Creek grades 9.7 % Pb and 4.1 % Zn, and only 0.5 % Cu. Gold grades for all of these smaller occurrences, as well as Arctic, range between undetectable to an average of about 1.2 g/t at the Smucker deposit (Newberry et al., 1997b). Silver grades locally reach 900 g/t in quartz-barite bands at Sun (Large). Metarhyolite samples from the Tom-Tom and Sun VMS occurrences in the Ambler district yielded ages similar to and slightly older than those at the Arctic deposit (McClelland et al., 2006).

**Bornite carbonate-hosted copper deposit**

Carbonate platform rocks exposed in a klippen a few kilometers south of the range front, host the undeveloped Bornite (or Ruby Creek) deposit (Fig. 6, Fig. 18), which is along Ruby Creek within the Ambler district (Runnels, 1969; Hitzman, 1986). The carbonate unit lies structurally above the Schist belt rocks (Till et al., 2008). The deposit is hosted in a 1000-m-thick Silurian to
Devonian marble and metadolostone sequence (Selby et al., 2009), the latter lithology perhaps of hydrothermal origin as it is highly brecciated, folded, and faulted. The breccia is syn-sedimentary and (or) hydrothermal. The lower greenschist facies carbonates are locally graphitic and phyllitic (Selby et al., 2009). Basin-margin faults appear to have localized the mineralization. Although middle Paleozoic mafic volcanic rocks and orthogneiss occur nearby, no igneous rocks are intimately associated with the Bornite deposit.

The **Bornite** deposit (Very Large), based on recent exploration by NovaCopper Inc., has an indicated open-pit resource of 14.1 Mt of 1.08 % Cu, and also an inferred resource of 109.6 Mt of 0.94 % Cu and 55.6 Mt of 2.81 % Cu for within pit and below pit ores, respectively (Davis et al., 2014), as well as highly anomalous Ag, Co, Ge, Ga, and Zn (Bernstein and Cox, 1986). Minerals associated with the copper ores include barite, bornite, carrollite, chalcocite, chalcopyrite, cymrite, galena, germanite, marcasite, pyrite, pyrrhotite, renierite, sphalerite, and tennantite-tetrahedrite. Cobalt is particularly anomalous, occurring within recrystallized pyrite rims and in carrollite (Bernstein and Cox, 1986). The ore is in stratiform zones in dolomite with stringers and veinlets, as well as the main breccia fillings. A two-stage paragenesis was described by Hitzman (1986) and Bernstein and Cox (1986). Stage 1 includes pyrite and some copper-bearing sulfides in veins, replacing matrix material in breccias (Fig. 20), and filling open-space in the dolostone. Alteration associated with stage 1 includes bleaching of phyllitic horizons, and development of hydrothermal dolomite and siderite in the limestone and phyllite, respectively. Stage 2, defining the main ore event, is characterized by copper minerals deposited in breccia matrix and in dolomite veins, with chalcopyrite replacing stage 1 massive pyrite, carrollite forming from the breakdown of cobalt-rich pyrite, and coarse-grained pyrobitumen associated with the copper minerals. The deposit is zoned from a central mineralized core of chalcopyrite, bornite, and chalcocite, outward to chalcopyrite and pyrite.

Hitzman (1986) favored ore formation at Bornite from basinal brines heated by magmatic activity, but stressed that the mineralization pre-dated the Jurassic-Cretaceous Brookian orogeny. Selby et al. (2009) dated the main stage sulfide event by Re-Os at ca. 374 Ma, an age that is similar to that of nearby VMS deposits. They identified a high Re concentration for the mineralization that was suggested to possibly reflect a role of organic matter in ore genesis. The coeval formation of VMS occurrences and the carbonate-hosted copper was suggested by Selby et al. (2009) to support the model of Hitzman et al. (1986) and Schmidt (1986) that regional extension and associated middle Paleozoic continental-margin back-arc magmatism was related to base metal deposition in the western Brooks Range. Davis et al. (2014) drew comparisons between the Bornite deposit and McArthur River (Australia), Tynagh (Ireland), Kipushi (Congo) and Tsumeb (Namibia).

**Ophiolitic chromite**

The allochthons of mafic and ultramafic rocks of the Angayucham terrane contain Jurassic occurrences of chromeite and platinum group elements (PGE). The more than 75 occurrences contain between 3 and 15 % chromite, are as long as 600

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**Figure 20.** Bornite mineralized carbonate breccia. Photo courtesy of Karen D. Kelley, USGS.
m, and were estimated by Foley et al. (1997) to contain between 0.64 Mt and 2.3 Mt of Cr₂O₃. The mineralization occurs in massive segregations, cumulate layers, and nodular aggregates in peridotite and dunite. Anomalous amounts of PGE are also recognized in some of the ultramafic rocks.

**Deposits in the eastern Brooks Range**
Economically less significant Cu, Mo, Pb, Sn, W, and Zn magmatic-hydrothermal occurrences are located in the eastern Brooks Range (e.g., Einaudi and Hitzman, 1986; Newberry et al., 1997a; Kurtak et al., 2002); none have been developed. Many of these are small Paleozoic porphyry and skarn occurrences that are overprinted and deformed during Mesozoic orogeny. Metaluminous, probably arc-related, ca. 400 Ma foliated porphyritic granite and granodiorite are cut by chalcopyrite-bearing stockworks (Newberry et al., 1986). Proximal Cu (Fig. 21A) and distal Pb-Zn skarns are associated with the porphyries where they intruded Devonian and older carbonates of the Arctic Alaska terrane (Newberry et al., 1997a). The Cu skarns (e.g., Chandalar) are Au-poor and contain garnet-magnetite and garnet-pyroxene-chalcopyrite assemblages. The Sn skarns (e.g., Okphilak) surround 380 Ma orthogneisses and are hosted by pre-Devonian carbonates. In contrast to this older mineralization, the early Tertiary Bear Mountain porphyry molybdenum occurrence is recognized at the eastern border of Alaska (Barker and Swainbank, 1985). Stockwork Mo-W mineralization is hosted in a breccia zone within a poorly exposed topaz-bearing rhyolite porphyry stock.

Small orogenic gold deposits are present in the Chandalar and Koyukuk districts in the southern Brooks Range. The districts historically yielded about 13.4 t of alluvial gold (Athey et al., 2014) and represent the only area across the entire Brooks Range with significant production. Placer production was greatest in the first decade of the 1900s, but working of the gravels continues today from modern stream deposits, bench placers, and older deep channels (e.g., Kurtak et al., 2002). Gold-bearing quartz veins, commonly containing abundant stibnite (Fig. 21B), have not been extensively developed and have yielded only about 0.5 t Au. The widely scattered veins fill high-angle faults and fractures that cut upper greenschist facies Devonian schist of the Arctic Alaska terrane. The ore-hosting structures likely formed during Albian regional uplift late during the Brooks Range orogeny (Dillon et al., 1987). Any genetic association with coeval Albian magmatism is, at most, indirect because the closest recognized granitic rocks of that age are 30-40 km south of the lode and placer occurrences.

**Western Alaska**

**Orogenic gold deposits and associated placers**
Small, typically poorly exposed, orogenic gold-bearing veins are widespread on the southwestern Seward Peninsula (Fig. 3; Mertie, 1918; Goldfarb et al., 1997). They mainly fill small fractures

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*Figure 21. Mineral occurrences in the eastern Brooks Range. A) Malachite-stained skarn outcrop at Hurricane-Diane, near Roberts Creek. B) Stibnite-gold veins exposed along a faulted contact on the south side of Sukakpak Mountain. Samples contained up to 65.4 g/t gold. Photos courtesy of Joseph M. Kurtak from BLM-Alaska Technical Report 50, July 2002.*
and joint sets within the calcareous metasiliceous unit of the Nome Complex, containing pure and impure marble, graphitic metasiliceous rock, pelitic schist, calc-schist, and mafic schist (Till et al., 2011, 2014). The only economic lode deposit was the **Big Hurrah** 70 km east of Nome; from 1903 to 1907, the mine produced slightly less than 1 t Au from veins in sheared metasedimentary and metavolcanic rocks, averaging 25 g/t Au (Cobb, 1978; Read and Meinert, 1986).

In 2008, a large open pit operation by NovaGold was initiated in the **Rock Creek** area (Fig. 3) about 10 km north of Nome in an attempt to produce gold from multiple veins in a bulk-tonnage type mining operation. The open pit was to have produced gold from an area with a high density of narrow sheeted quartz veins hosted by green-schist facies schist, quartzite, and marble (Otto et al., 2009), part of the calcareous siliceous unit of Till et al. (2014). The combined probable + indicated + inferred reserves were stated as 18 t Au at a 0.6 g/t cutoff, with an average grade just above 1 g/t Au. In addition, the mill at Rock Creek was planned to handle higher grade ores of 3-5 g/t to be mined by and trucked from a new open pit at the historic Big Hurrah deposit, with estimated remaining reserves of 9 t Au. However, after an initial gold pour in fall of 2008, all lode mining activities ceased due to failure of mill equipment and environmental issues.

The Seward Peninsula orogenic gold province probably extends into eastern Russia, where large deposits such as Mayskoye and Karalveem are hosted by Middle Triassic sedimentary rocks and late Aptian to early Albian granite and granodiorite (Goldfarb et al., 2014). Poorly constrained ages of metamorphism and ore formation suggest a correlation of the gold event across the Bering Sea. In eastern Russia, the orogenic deposits are associated with regional anticlines in the Chukotka part of the Arctic Alaska-Chukotka microplate. Large-scale structures are better developed in Chukotka than the Seward Peninsula, which might explain the much greater lode gold favorability.

The areas of orogenic gold veins lack known igneous intrusive rocks, although potassic granite dated between 100 and 85 Ma (Amato and Wright, 1997; Till et al., 2011) is common 40-50 km to the north and east of the gold deposits. The rocks of the Nome Complex underwent blueschist metamorphism before 120 Ma. During the Albian or early Late Cretaceous (Amato et al., 1994), rocks of the Nome Complex were overprinted by a Barrovian metamorphic event capable of producing a large volume of auriferous metamorphic fluid (e.g., Read and Meinert, 1986). Metamorphic grade locally reached upper amphibolite to granulite facies during this event (Till et al., 2011). The absence of a major, deep-crustal regional structure cutting the area affected by the mid-Cretaceous metamorphism may explain formation of hundreds of small gold-bearing veins scattered throughout the schist, rather than a small number of large tonnage ore deposits.

Placer gold is widespread throughout the south-western side of the Seward Peninsula (Fig. 10A), present in alluvial, colluvial, glacial, and particularly marine strandline deposits (Cobb, 1974). Active stream channels, as well as benches with old alluvial or glacial channels high along stream walls, were productive, and in places they yielded large nuggets (e.g., Moffit, 1913). However, the majority of the recovered gold was from **beach deposits of the Nome area** (Large). The first gold discovery was on the present-day beaches of Nome, which yielded an estimated 3-4 t of Au along 60 km of coastline (Fig. 10B). Soon after this initial discovery on the “first beach”, it was realized that the bulk of the gold was located slightly inland within ancient beach deposits, and these older deposits were responsible for most of the recovered 155 t Au. Six ancient beach marine platforms were located above present sea-level and an equal number were located below present sea-level (Nelson and Hopkins,
The second beach was located about 12 m above sea-level and the third beach about 26 m above. The gold within the third beach was located just above bedrock, in beach sands and river gravels at the bottom of 10-15 m deep shafts (Mertie, 1913). The most landward submarine beach was discovered 400-500 m inland and 6-7 m below present sea-level. Metz (1978) estimated reserves of 37 t Au remaining after mining, mainly in the second, third, submarine, and Monroeville beaches. Large-scale mining of the alluvial gold ceased in 1962 (Cobb, 1974), although recreational mining of many of the beaches continues today.

The marine benches formed in the late Pliocene to Pleistocene as a result of relative sea-level fluctuations. The gold-bearing gravels in the benches were deposited by glaciers on top of the schist bedrock and the fine-grained marine sediments. The gold in the till was then reworked and concentrated by both fluvial and marine processes. The latter included littoral currents and waves, bottom currents, and shoreline migration (Kaufman and Hopkins, 1989). Offshore alluvial gold occurrences are also abundant for about 15 km in length parallel to the present beach at Nome, for distances of about 5 km outward into the Bering Sea and to depths of 20 m (Nelson and Hopkins, 1972; Kaufman and Hopkins, 1989). The greatest concentration of gold is in the upper 4 m of seafloor sediment. These concentrations were also products of Pleistocene glaciation that carried eroded lode, alluvial, and beach placers offshore. Nelson and Hopkins (1972) indicated the offshore lag gravels contained, on average, eight times the amount of gold as the parent till.

The Lost River tin skarn deposit in the York Mountains, about 135 km northwest of Nome, produced 314 t Sn in the first half of the 1950s (Lorain et al., 1958) and is the largest historic tin producer in Alaska. The mineralization is mostly adjacent to the buried cupola of one of a series of ca. 80 Ma evolved granite intrusions in the western Seward Peninsula that intrude Early Ordovician limestone. Four stages of hydrothermal activity have been described (Dobson, 1982; Hudson and Reed, 1997; Newberry et al., 1997a). Initial Mg-Al-Fe-rich skarn formation included some tin enrichment in silicate phases, such as garnet and idocrase. An overprinting hydroskarn-forming event included deposition of cassiterite, as well as pyrite, pyrrhotite, chalcopyrite, sphalerite, scheelite, fluorite, tourmaline, biotite, and hornblende. The tin-rich mineralization is associated with greisenization along the margin of the intrusion. This was followed by formation of post-skarn fluorite-white mica veins, and then solution breccias that overprinted and oxidized many of the ore minerals and deposited fluorite and kaolinite. Each stage may have further concentrated the tin and upgraded the ore deposit (Dobson, 1982).

The past production at Lost River, from ore that averaged 1.42 % Sn, was from the so-called Cassiterite Dike, which occurs as a muscovite-quartz-tourmaline-topaz roof greisen replacing quartz porphyry dikes and limestone at the top of the buried intrusion (Sainsbury, 1964, 1988; Hudson and Reed, 1997). The greisen is as much as 30-60 % muscovite (Dobson, 1982). The greisenization resulted in deposition of cassiterite and sulfide minerals within the altered dikes that have a maximum width of about 6 m. The metallic minerals are disseminated and in cross-cutting veins in the greisen. A clay alteration overprint has commonly broken down all mineral grains except for the cassiterite. Zoning of
a beryllium-rich assemblage, including fluorite, chrysoberyl, beryllium diaspore, white mica, tourmaline, and minor beryl, euclase, bertrandite, and phenacite is noted both outward and upward from the cassiterite-rich assemblage. Reported resource estimates for the tin skarn are approximately 55,000 t Sn from ore with an average grade of 0.26 % Sn (Hudson and Reed, 1997).

The other major lode tin deposit is the undeveloped Kougarok deposit, with estimated reserves of 90,000 t Sn. The deposit was discovered based on a stream sediment survey in the late 1970s (Puchner, 1986). The cassiterite is associated with zinnwaldite and muscovite in a greisened late stage granite, similar to Lost River. Relatively less abundant cassiterite is associated with disseminations and stringers of tourmaline, axinite, chlorite, and quartz in altered schist country rocks.

Almost all of the tin mined in Alaska has been from placer occurrences of the Seward Peninsula, particularly along Cape Creek draining Cape Mountain (Fig. 3; 1676 t Sn) and from Buck Creek and upper tributaries to the Anikovik Rover near Potato Mountain (1000 t Sn). Production was discontinuous between 1911 and 1990 (Hudson and Reed, 1997). The alluvial tin in these areas is sourced from the erosion of upstream cassiterite-bearing quartz veins and skarns in deeply eroded granites (e.g., Swanson et al., 1990). Hudson and Reed (1997) indicate the favorable pay streaks may reach a few kilometers in maximum length, are as much as 100 m in width, and have variable thickness to a maximum of a few tens of meters.

Other deposit types of the Seward Peninsula

Metacarbonate and metasomatic rocks of the Nome Complex in the southern half of the Seward Peninsula also host small, but widespread deformed and metamorphosed Zn-Pb±Ag deposits and occurrences (Fig. 3; Slack et al., 2014). These occur predominantly in the same rocks and over much of the same area as the orogenic gold deposits, and have no spatially associated igneous rocks. In contrast to the Cretaceous gold deposits, however, these base metal deposits (e.g., Aurora Creek, Galena, Nelson, Quarry, Wheeler North) are present as pre-Carboniferous sub-seafloor replacement (clastic-dominated Pb-Zn) deposits (Fig. 22) that have had only sporadic, small-scale development. The sphalerite, galena, and pyrite ores occur as lenses of disseminated to semi-massive sulfide, within quartz, siderite, and ankerite gangue, and have local concentrations of barite and fluorite. Slack et al. (2014) pointed out that sphalerite aligned in foliations and folded barite bodies are consistent with deformation of the occurrences during the Brooks Range orogeny.

A second group of Zn-Pb-Ag occurrences (e.g., Hannum, Independence, Omilak, Foster) is located in central Seward Peninsula near the Cretaceous intrusions in amphibolite facies rocks of the Nome Group. The sulfides occur as disseminations, and in veins, pods, and stringers that replace the schists and marble and are present along the contacts between the units. Because the occurrences appear undeformed and unmetamorphosed, and have very high \(^{206}\text{Pb}/^{204}\text{Pb}\) ratios (Ayuso et al., 2014), they are ap-
parently no older than Cretaceous. Furthermore, the spatial association with the granitic rocks hints at a genetic association.

The Boulder Creek or Death Valley uranium deposit in eastern Seward Peninsula (Fig. 3), described by Dickinson et al. (1987), is the northernmost sandstone-type uranium deposit identified in the world. Discovered in 1977, it is hosted in early Eocene continental arkosic and conglomeritic sedimentary rocks that unconformably overlie a weathered Cretaceous granite. The mineralization is about 100 m thick and consists of coffinite and minor pyrite; in places of supergene enrichment, the uranium minerals have been converted to meta-autunite. The Boulder Creek deposit is estimated to contain 453 t U\(_3\)O\(_8\) with an average grade of 0.27 % U\(_3\)O\(_8\); the deposit has been explored, but not developed. The deposit formed when uranium was leached from the weathered granite by oxidized groundwater and deposited at a redox front represented by carbonaceous matter and coal within elastic sediments. A much more temperate or subtropical climate would have been required in the early Eocene to form such a deposit.

Ruby geanticline
Southeast of the Seward Peninsula, the relatively isolated Illinois Creek polymetallic vein
deposit formed along the southwestern side of the Ruby Geanticline (Ruby terrane of figure 6), a regional NE-trending uplift of mainly early Paleozoic metasedimentary rocks of mainly continental-margin affinity in interior Alaska. Patton et al. (1994) correlated these rocks with those of the southern Brooks Range, although their history remains unclear (e.g., Bradley et al., 2007). The Illinois Creek deposit (located in Ruby-Poorman district; see below figure 23), discovered in the Kaiyuh Hills in 1980, was mined from a seasonal open-pit heap leach operation between 1997 and 2004, producing about 5 t Au and 24 t Ag from ore averaging 2 g/t Au and 43 g/t Ag. At least equal amounts of ore remained but mining ceased due to financial and environmental issues. The mined out oxide ore was hosted in a shear zone that extends for at least 2 km within a pelitic schist; the nearest intrusion is 12 km to the north of the deposit (Flannigan, 1998). Gold occurs in oxidized pyrite grains within Fe-stained vein quartz, but at depth it occurs with numerous polymetallic sulfides and sulfosalt minerals. The ca. 113 Ma age for both the Illinois Creek deposit and the granite to the north led Flannigan (1998) to classify the deposit as an intrusion-related vein system. Recent exploration in the area has focused on polymetallic replacement and porphyry copper potential, but this part of west-central Alaska is extremely remote and much of the area is under cover so the resource potential of the Ruby Geanticline remains unclear.

**Metallogeny of the Farewell terrane**

The Farewell terrane (Fig. 4), connected with Arctic Alaska terrane into the Devonian, has resource favorability for deposits similar to those of the southwestern part of the Brooks Range. Schmidt (1997) suggested that some of the early to middle Paleozoic rocks of the Farewell terrane have potential for undiscovered clastic-dominated Pb-Zn and barite deposits. The Reef Ridge zinc deposit (420,000 t grading 17.4 % Zn; Fig. 4), as well as other nearby smaller occurrences, are hosted by Early to Middle Devonian shallow water platform dolomite. They are dominated by smithsonite that formed during cold, humid weathering of sphalerite and carbonates in the late Tertiary and Holocene (Santoro et al., 2015). The precursor sulfide ores were interpreted as Mississippi Valley-type deposits by Schmidt (1997). However, as pointed out by Santoro et al. (2015), a spatial association with Late Cretaceous porphyry prospects and the Nixon Fork Cu-Au skarn to the southwest (see below discussion of SW Alaska) suggests the Reef Ridge deposit and related Zn-nonsulfide replacement and veinlet occurrences formed from supergene alteration of Late Cretaceous magmatic hydrothermal vein deposits. In addition, in many parts of the Farewell terrane there are relatively unexplored occurrences of barite and phosphates (Bundtzen and Gilbert, 1991; Dumoulin, 2015).

**Eastern interior Alaska**

**Gold metallogeny**

Eastern interior Alaska contains significant gold resources in orogenic gold deposits, a world-class intrusion-related gold deposit, and widespread alluvial deposits. The deposits define a part of what has become widely known to explorationists as the Tintina Gold Belt or Tintina Gold Province (Fig. 23), which extends for >1500 km from southwestern Alaska, through central and eastern Alaska, and into western Yukon (e.g., Goldfarb et al., 2000; Tucker and Smith, 2000). The belt, however, consists of many different gold deposit types of various ages and therefore the term Tintina Gold Belt is more of a promotional term, lacking scientific merit, rather than one reflecting a well-defined belt of related auriferous ore deposits. In the Fairbanks district, the approximately 300 t Au Fort Knox deposit (Very Large) (Fig. 24A) is perhaps the only global example of an economic reduced intrusion-related gold system; the term “system” is used by many workers due to zoning of deposit types around a causative intrusion (e.g., Hart et al., 2002). These systems were first described as including potential economic gold deposits by Thompson et al. (1999), Thompson and Newberry (2000), and Lang et al. (2000). This model was predominantly based upon the Fort Knox deposit in Alaska, and similar prospects, such as Scheelite Dome, Dublin Gulch, and Clear Creek (Fig. 23), in adjacent Yukon. Prior to the development of the reduced intrusion-related gold system model, similar magmatic and mineralogic features had previously been described for gold skarns related to reduced intrusions (Meinert, 1989; 1995).

The Fort Knox deposit, about 25 km northeast
of the city of Fairbanks, is located in the apex of the variably porphyritic, moderately reduced, monzogranite to granodiorite Vogt stock (Bakke, 1995). The 92.5 Ma host stock, which intrudes the Proterozoic to middle Paleozoic Fairbanks schist, is characterized by a low magnetic susceptibility and an $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio of 0.15-0.30 (Hart et al., 2004a). The stock had a pre-mining surface exposure of 1100 x 600 m. Bakke (1995) subdivided the stock into an early biotite-rich fine-grained granite, a medium-grained porphyritic granite, and a youngest coarse-grained seriate porphyritic granite. Isotopic data from the ore-hosting stock at Fort Knox indicate a magma origin reflecting both mantle and crustal contributions (Haynes and et al., 2005).

The gold occurs in steeply-dipping, commonly sheeted, quartz-K-feldspar veins, and in planar quartz veins that occur along later gently- to moderately-dipping shear zones cutting the igneous rocks (Fig. 24B-C). The sheeted veins have been referred to as pegmatites by some workers, but are definitely veins with clear to gray quartz and K-feldspar grains that were deposited by a hydrothermal fluid. The density of the sheeted veins strongly controls the ore grade. The veins generally fill northwest-striking, shallowly to moderately southwest dipping shear zones, and individual veins range in width from 0.3-1.5 m. Some of the veining also may appear as thin quartz stockworks. High fineness gold in the veins is commonly intergrown with native bismuth, bismuthinite, and tellurobismuth (McCoy et al., 1997). Total sulfide volume is typically much less than 1 % and bismuthinite is commonly the most abundant sulfide phase in the veins. Other minor sulfides include pyrite, pyrrhotite, arsenopyrite, and molybdenite; traces of scheelite are also present. Alteration phases include K-feldspar, albite, biotite, sericite, and ankerite; they generally define haloes surrounding veins of only a few centimeters (Fig. 24B). A Re-Os date on hydrothermal molybdenite of 92.4±1.2 Ma is identical to the crystallization age of the Fort Knox stock (Selby et al., 2002).

A set of late, through-going, northeast-trending shear veins cut the stock and sheeted veins (Fig. 24D). Their structural evolution has not been well studied, but they are important in adding grade to the total resource and increasing the overall grade of the historically recovered ores.
to about 0.9 g/t Au (Arne Bakke, oral commun., 2003). It is likely that if it were not for these auriferous late shears, the Fort Knox gold deposit would not have been economic.

Numerous smaller gold deposits are also distributed throughout the Fairbanks district (Fig. 25; Goldfarb et al., 1997; McCoy et al., 1997, Bakke et al., 2000; Hart et al., 2002). The True North deposit (4.7 t Au produced), originally prospected 100 years earlier as an antimony prospect, was briefly mined for gold, providing higher-grade feed for the Fort Knox mill, from 2001-2005. This deposit, located about 20 km northwest of Fort Knox, is comprised of gold in quartz veinlets, disseminations, and breccia along a shallow thrust in carbonaceous felsic schist. The fine-grained gold is associated with pyrite, arsenopyrite, and stibnite. Igneous rocks are not recognized at the deposit. The Gil deposit, about 10 km east of Fort Knox, is a sheeted gold-bearing vein system in calc-silicates that likely formed above an as yet unroofed pluton intruding schist; other small and subeconomic tungsten skarns also occur in the same area. Many shear-hosted gold-bearing quartz veins, typically dominated by arsenopyrite or stibnite, occur throughout the district. The largest, the Ryan Lode (Medium) (<1 t Au produced), occurs along the sheared margin of a tonalite stock. The majority of the schist-hosted deposits (e.g., Grant, Cleary Hill, Newsboy, Hi-Yu, Christina, Tolvana) are along steeply-dipping, NE- or NW-trending fault zones and have characteristic quartz-sericite-ankerite alteration haloes. $^{40}$Ar/$^{39}$Ar hydrothermal mica dates of 103 Ma for the Hi-Yu deposit (Goldfarb, unpub. data) and 89 Ma for the Ryan Lode (McCoy et al., 1997), as well as a Re-Os sulfide date of 92 Ma for Tolovana (Goldfarb, unpub. data), suggest that the schist-hosted deposits both overlap and pre-date Cretaceous magmatism in the district. Many of these deposits hosted in the Fairbanks Schist are best classified as orogenic gold deposits, although the Gil deposit may be similar to Fort Knox in having a close association with magmatic-hydrothermal activity.

Erosion of the widespread auriferous quartz veins has yielded alluvial concentrations responsible for 257 t Au production from placer deposits of the Fairbanks district (Very Large) (Fig. 11). Production peaked during the first few years of mining after discovery of alluvial concentrations in 1902 and during a lengthy period of dredging between 1928 and 1963. Most production came from the watersheds of Cleary, Fairbanks, Ester,
Dome, and Goldstream Creeks. The most productive pay streaks were typically less than 100 m in width and at or near the bedrock contacts (Tuck, 1968). The late Pliocene to middle Pleistocene Cripple and Fox Gravels were most productive (Pewe, 1975). Metz (1991) suggested the most important placers formed in streams forming a barbed drainage pattern, where tributaries flow in the opposite direction to their master streams and thus join these streams in a hook-shaped bend. Such drainages developed in areas of basement structures, where channels were captured and altered by the basement features.

The Goodpaster mining district, located about 140 km southeast of Fairbanks (Fig. 23), was characterized by only very minor gold production from small vein and placer occurrences for 100 years. However, high-grade (avg. 12.5 g/t Au), shear-hosted veins cutting Proterozoic to middle Paleozoic biotite-quartz-feldspar or thogneiss and paragneiss of the Yukon-Tanana terrane (Day et al., 2003) were discovered at Pogo (Large) in 1996. Underground mining began ten years later. As of 2015, Pogo is the largest gold producing mine in Alaska, producing about 11 t Au/year and with a present production, reserve, and resource total of 220 t Au.

The gold-bearing veins at Pogo, termed the Liese vein system (Fig. 26), occur as three individual, laminated, stacked veins that dip shallowly to the northwest. The ductile to brittle veins average 7 m in thickness, although they are locally as thick as 30 m, and have an aerial extent of 1.4 x 0.7 km (Smith et al., 1999; Rhys et al., 2003). The largest vein has a down-dip extent of >1.7 km. The dominant foliation and stretching lineations at the Pogo deposit are interpreted to be consistent with a regional compressive regime forming the main flat veins and a switch over to extensional stresses to form steeper brittle tensional veins that are also gold-bearing (Eric Jensen, University of Alaska-Fairbanks, written commun., 2009). Sulfide phases, comprising about 3% of the veins, include arsenopyrite, pyrite, pyrrhotite, loellingite, chalcopyrite, and molybdenite; Bi- and Te-bearing tellurides are also present. Alteration phases include biotite, quartz, sericite, K-feldspar, ferroan dolomite, and chlorite.

Reduced granites and tonalites of the Goodpaster batholith are located a few kilometers north of Pogo. These rocks were intruded ca. 109-103 Ma, during the final stages of regional metamorphism and deformation (Dilworth et al., 2007). In contrast to those associated with the Fort Knox deposit that have a significant mantle component, the magmas that formed the Goodpaster batholith are mainly crustal melts. Igneous activity overlaps with the 104.2 Ma mineralization age as determined by Re-Os analysis of ore-related molybdenite (Selby et al., 2002). The deposit is cut by a post-kinematic 94 Ma diorite. The
temporal association of the intrusions with the gold event has led most workers to define Pogo as an intrusion-related gold system. However, many features of the deposit suggest that it is no different than most orogenic gold deposits and a genetic relationship to magmatism is far from conclusive (Goldfarb et al., 2000, 2005).

The Tolovana mining district is located 75 km northwest of Fairbanks (Fig. 23). It has yielded about 16 t of placer gold in the past 100 years, and significant alluvial resources are still present. Small sulfide- and gold-bearing quartz veins in Paleozoic schist, volcanic rocks, and ultramafic rocks have been prospected on Money Knob, a domal feature in the headwaters of many of the gold-bearing creeks. A bulk minable target was defined in the area of the Money Knob prospects in 2007, namely the Livengood deposit (Very Large), with present measured, indicated, and inferred resource estimates of about 570 t Au at an 0.3 g/t Au cutoff and averaging just above 0.5 g/t Au (International Tower Hill website).

There is little detailed information on the Livengood deposit, with the only published geology in Pontius et al. (2010). The mineralization seems to be within a complex series of fault slices of Cambrian mafic and ultramafic rocks and Devonian clastic and volcanic rocks, with most ore in the latter sequence. The mineralization at Money Knob is both disseminated and vein-hosted. Arsenopyrite, pyrite, and lesser stibnite are commonly associated with the gold. Alteration phases are dominantly biotite, sericite, and albite, although a silica-carbonate-talc assemblage is reported in the ultramafic rocks. Mid-Cretaceous stocks, dikes, and sills are widespread in the area of Money Knob, leading most reports on the deposit to define it as an intrusion-related gold system, although descriptions of some of the mineralization styles on the dome also are consistent with an orogenic gold deposit classification. The lack of anomalous Bi and Te was argued by Pontius et al. (2010) to indicate that Money Knob was a “distal” intrusion-related gold system.

Important lode gold deposits have yet to be discovered in other placer mining districts of east-central Alaska, such as Circle, Rampart, Fortymile, and Hot Springs. Yeend (1991) noted that in the Circle district, alluvial gold was mainly concentrated in the lower 1 m of stream gravels and the underlying upper 0.5 m of bed-
A strong correlation of gold-rich channels with mafic schist was observed, but significant lode sources have not been located in more than 100 years of prospecting in the district. Similarly, in the Fortymile district, the gravel-bedrock interface has localized most of the recovered alluvial gold (Yeend, 1996). It remains uncertain as to whether (1) all gold-bearing lodes are small and scattered, (2) most lode endowment has been lost to erosion, and (or) (3) future lode discoveries, such as in the Fairbanks, Tolovana, and Goodpaster districts, are still possible in these other historic placer districts. Allan et al. (2013) have noted that a few small gold-bearing veins in the Fortymile district could have formed at any time from Middle Jurassic to early Tertiary, so therefore even the ages of potential source lodes for the alluvial gold are not clear.

**Volcanogenic massive sulfide deposits**

Poly metallic VMS deposits of the Bonnifield and Delta districts are located along the north side of the Denali fault system in the northern foothills of the eastern Alaska Range (Fig. 6). The deposits are part of a series of Zn-Pb-Ag-Au±Cu massive sulfides formed during back-arc extension and the Devonian-Mississippian opening of the Slide Mountain Ocean (Nelson et al., 2013). Tertiary dextral displacement has resulted in wide distribution of these deposits in western Canada and eastern Alaska. The northernmost deposits hosted in the parautochthonous rocks of Yukon-Tanana terrane (e.g., Bonnifield and Delta districts) have had no development.

The Bonnifield district, about 80 km south of Fairbanks, hosts 26 siliclastic-felsic type VMS occurrences in peralkaline metarhyolite and intercalated graphitic and siliceous argillite of the Totatlanika Schist (Fig. 27; Dusel-Bacon et al., 2012). The ca. 373-356 Ma mineralization is primarily pyrite and sphalerite, with lesser galena, tetrahedrite, and chalcopyrite (Fig. 28). High-grade zones of sulfide minerals are locally as thick as 13 m. The largest reported resource is for the combined Dry Creek (Red Mountain) and WTF deposits, with 5.7 Mt at 4.4-6 % Zn, 2.5 % Pb, 0.3 % Cu, 94-179 g/t Ag, and 0.5-0.9 g/t Au. The massive sulfide bodies have been overprinted by Mesozoic deformation and greenschist facies metamorphism. Similar mineralization of the same age, although with smaller recognized resources, is characteristic of the Delta
district about 150 km southeast of the Bonnifield deposits (Fig. 6; Dashevsky et al., 2003). Globally, such VMS occurrences associated with alkalic magmatism have to a large degree proven to be uneconomic (Jim Franklin, oral commun., 2015).

**Porphyry copper deposits**

A number of porphyry copper deposits are located within the high elevations of the eastern Alaska Range (Fig. 5, 29, 30A-C) to the south of the Denali fault system, but their remoteness and low metal grades has long discouraged development. Intrusions hosting the Cu±Mo-Au stockworks and disseminations (Fig. 30D-F) were emplaced at ca. 117-105 Ma and are exposed over an area of about 250-300 km² (Richter et al., 1976). They intrude late Paleozoic to Triassic metasedimentary and metavolcanic rocks of the Wrangellia composite terrane close to its landward contact with the Jurassic-Cretaceous flysch of the Nutzotin basin. The porphyry deposits probably formed along what was then the continental margin, more than 1,000 km south of their present-day latitude, during the cessation of basin sedimentation and the onset of a continental margin transpressional regime (Fig. 31). The giant Pebble porphyry deposit in southwestern Alaska (see below) may be part of the same middle Cretaceous post-subduction magmatic-hydrothermal episode (Goldfarb et al., 2013).

The mineralized intrusions in the eastern Alaska Range were estimated to cumulatively contain about 1,000 Mt averaging 0.25-0.35 % Cu (Richter et al., 1975). The Klein Creek pluton consists of granodiorite, monzonite, and diorite and hosts the Baultoff, Carl Creek, and Horsfeld deposits along the Alaska-Yukon boundary. About 50 km to the southwest, the major part of the inferred resource is present in the Orange Hill and Bond Creek deposits that are hosted in the granodiorite and quartz diorite of the Nebesna batholith. The **Orange Hill** deposit (Very Large)
Figure 30. Copper porphyry occurrences in the eastern Alaska Range. A) View on top of Orange Hill porphyry looking N to NNE. B) Bond Creek Ridge, site of Bond Creek porphyry. C) Pyritic talus slope at the Baultoff deposit. D) Low-grade stockwork veining in the Orange Hill intrusive in the upper reaches of California Gulch. E) Drill core showing quartz-chalcopyrite vein from Orange Hill. F) Stockwork quartz-chalcopyrite veining in heavily altered volcanic rocks, Bond Creek. Photos courtesy of Garth Graham USGS.
Fig. 31. Palinspastic reconstruction of western North America from the Mesozoic to the Tertiary (generalized from Rea and Dixon, 1983; Umhoefer, 2003; and Smith, 2007). A) One possible scenario for Farallon-Pacific plate spreading during the poorly understood ca. 119-83 Ma period of the Cretaceous Normal Superchron. At some time between 119 and 83 Ma, the Kula plate likely separated from the Farallon plate. Porphyry-related intrusions are dated at ca. 100-90 Ma to the north and at 115-105 Ma to the south along the boundaries of the flysch basins with Wrangellia and Peninsular terranes. The Pebble deposit was the most northerly of the belt of porphyry deposits formed within the diachronously closing Kahiltna-Nutzotin Ocean and then translated along the continental margin. B-E Interpreted time slices of preferred important far-field stresses leading to formation and translation of the middle Cretaceous porphyry deposits. B) Early Cretaceous sinistral translation of Wrangellia composite terrane along the margin of North America. C) Mid-Cretaceous Mendocino fracture zone (fz) splitting Farallon plate into north and south entities. Dextral motion along margin. D) Late Cretaceous Chugach sediments shed into subduction zone. Kula plate now being subducted beneath Wrangellia. Continued dextral translation along margin. E) Tertiary continued buildup of Chugach accretionary complex. Oroclinal bending of Alaska with continued translation of Wrangellia. After Goldfarb et al. (2013).
contains 100–320 Mt grading 0.30–0.35 % Cu and 0.02–0.03 % Mo. The Bond Creek deposit (Very Large) contains an estimated 500 Mt grading 0.15–0.40 % Cu and 0.02 % Mo (Young et al., 1997). In addition, Orange Hill is estimated to contain about 45 t Au at 0.15 g/t (Richter et al., 1975). The intrusive complex at Nadesna is associated with typical potassic, phyllic, argillic, and propylitic alteration assemblages, whereas that at Klein Creek only shows potassic and propylitic phases (Hollister et al., 1975; Hollister, 1978). The deposits were included in the Wrangell-St. Elias National Park and Preserve established in 1980 and are thus not exploration targets. However, west of the park, about 100-150 km from Orange Hill, the large Ahtell pluton has been the site of recent exploration for middle Cretaceous Cu-Mo-Au porphyry mineralization (Myers et al., 2010).

The Nadesna gold-rich copper skarn is located about 20 km northwest of the Orange Hill porphyry (Fig. 29). A middle Cretaceous pluton of quartz diorite and granodiorite, similar to rocks of the Nadesna batholith, intrudes Late Triassic Chitistone Limestone (Wayland, 1943; Newberry, 1986). The calcic skarn is dominated by chalcopyrite and garnet, with calc-silicate zoning from garnet outward to pyroxene and then unaltered marble (Newberry et al., 1997a). Minor amounts of gold were produced in the first half of the 20th century from pyrite veins cutting the marble and from sulfide-rich skarn.

South-central Alaska

**Kennecott copper deposits**

The Kennecott copper deposits are located in the southern part of the Wrangell Mountains about 80 km south of the Orange Hill and Bond Creek porphyry deposits and the historic workings are also now within the boundaries of the Wrangell-St. Elias National Park and Preserve (Fig. 29). The copper and silver ores, which were mined from five major deposits (medium sized) (e.g., Bonanza, Jumbo, Mother Lode, Erie, and Glacier), are hosted within steep faults in the base of the Late Triassic Chitistone Limestone of the Wrangellia terrane. They all occur within about 100 m of the underlying subaerial basalt of the Triassic Nikolai Greenstone (Fig. 32A; MacKevett et al., 1997). The high-grade deposits averaged about 13 % copper and were dominated by a chalcocite-djurleite assemblage (Fig. 32B). MacKevett et al. (1997) described a paragenesis where early pyrite was replaced by chalcopyrite, then bornite, and ultimately by the chalcocite and djurleite.

Several different models exist for ore genesis. MacKevett et al. (1997) argued that the ore-hosting faults in the limestone formed during the Jurassic-Cretaceous collision of Wrangellia...
with North America. Residual brines that were trapped in sedimentary rocks descended into the basalt with a high background concentration of 160 ppm Cu and leached copper. Deposition of sulfides occurred as fluids were focused into the faults in the overlying limestone. As temperatures waned during the hydrothermal event, chalcocite and djurleite replaced the earlier formed bornite. In contrast, Silberman et al. (1978) argued that this final replacement of the bornite was a distinct event that happened in the late Tertiary as young magmatic events facilitated the circulation of meteoric fluids. Silberman et al. (1980) also suggested that a K-Ar age of 112±11 Ma for chloritized greenstone stratigraphically below the main ores represents the age of deposition for the bornite.

**Gold metallogeny**

Small orogenic gold deposits are scattered through the Chugach and Kenai Mountains and are hosted by flysch of the Valdez Group (Fig. 33A-D). Many occur as discordant quartz veins in saddle reefs (Fig. 33C), joint sets, tensional fractures, and sheared margins to slightly older intrusions within districts such as Port Valdez, Port Wells, Girdwood, Hope-Sunrise, Moose Pass, and Nuka Bay (Fig. 33E; Goldfarb et al., 1986). High-grade veins are typically characterized by abundant wallrock ribbons and, locally, auriferous brecciated veins are important. Arsenopyrite is the most common sulfide phase and coarse visible gold is typical of many deposits. There is little visible wallrock alteration, reflecting both the small size of the deposits and fluids in thermal equilibrium with metasedimentary country rocks (Goldfarb et al., 1986). Most gold deposits in the Chugach and Kenai Mountains are spatially associated with medium greenschist facies Valdez Group rocks, which are characterized by incipient metamorphic biotite. The age of mineralization throughout this part of the accretionary complex is ca. 57-53 Ma (Goldfarb et al., 1997) and the hydrothermal event may be attributed to subduction of a spreading oceanic ridge (Fig. 33F; Haeussler et al., 1995; Bradley et al., 2003).

Orogenic gold deposits of the Willow Creek district are located a few tens of kilometers north of
the Castle Mountain strand of the Border Ranges fault system and within the Early to Middle Jurassic Talkeetna arc developed in the Wrangellia composite terrane. The mines in the district (Fig. 34A; e.g., Independence, Lucky Shot, Gold Bullion) defined the largest lode gold producing region north of latitude 60 degrees prior to the development in the past two decades of the Fort Knox and Pogo deposits near Fairbanks. The Willow Creek mines also yielded minor scheelite during World War II. Most of the dozens of gold occurrences are present as auriferous quartz veins in shear zones cutting the southern margin of the Late Cretaceous-early Tertiary Talkeetna Mountains granodiorite batholith that was emplaced into the Jurassic arc (Ray, 1954; Madden-McGuire et al., 1989). Veins show evidence of crack-seal events (Fig. 34B) and more brittle brecciation events, with pyrite as the dominant sulfide, along with gold, tellurides, scheelite, and locally abundant base metal-bearing sulfides and sulfosalts.

The plutonic rocks of the Talkeetna Mountains batholith in the area of the Willow Creek gold deposits were emplaced between ca. 74-67 Ma.
Most of the gold-bearing veins likely formed at ca. 66 Ma, as the just crystallized batholith was being rapidly uplifted (Steve Harlan, U.S. National Science Foundation, unpub. data). The Castle Mountain fault (Fig. 34A), exposed about 5 km south of the gold deposits and along the edge of the batholith, is estimated to have undergone 130 km of slip post-65 Ma (Pavlis and Roeske, 2007), which would have begun at approximately the time of gold deposition. Such strike-slip motion is commonly associated with regional uplift and formation of orogenic gold ores in a retrograde P-T environment of an evolving orogen (Goldfarb et al., 2005).

In the case of the Willow Creek gold district, orogenic gold deposits formed within the margin of the subduction-related batholith itself and not in metamorphosed fore-arc or back-arc regions, as is characteristic of most North American Cordilleran settings (e.g., Goldfarb et al., 2008).

**Volcanogenic massive sulfides**

Paleocene and Eocene metasedimentary rocks and greenstones of the Prince William terrane host massive to semi-massive sulfide occurrences that were developed and mined from 1897-1930 along the fjord margins and on islands of Prince William Sound (A-C). The largest body, producing about 95% of the copper (83,000 t Cu) from the region, from both open pit and underground workings, was the 300 m x 120 m Beatson massive sulfide (Fig. 35A) on LaTouche Island (Bateman, 1924). After the Kennecott deposits, Beatson was the second largest copper producer in Alaska, although mainly economic because of its easy tidewater access and the simultaneous mining of the Kenneecott orebodies inland to the north. The numerous VMS occurrences, dominated by pyrrhotite, pyrite, chalcopyrite, and sphalerite (Moffit and Fellows, 1950), are typically localized in sediments deposited along the Kula-Resurrection Ridge that was spreading near the edge of the North American continent (e.g., Bradley et al., 2003). The seafloor volcanism occurred at ca. 57 Ma, shortly before the host rocks for the occurrences were accreted to the continent (Crowe et al., 1992).

**Chromite**

A belt of dunite and pyroxenite with chromite pods, wisps, bands, and disseminations is discontinuously exposed for more than 1,000 km along the Border Ranges fault system, separating the Chugach and Wrangellia terranes (Guild, 1942; Foley et al., 1997). The largest deposit within the Border Ranges Ultramafic-Mafic Complex at Red Mountain (located about 30 km west of Nuka Bay mine on figure 33E) with 1.4 Mt of 5-6 % Cr₂O₃ yielded small amounts of chromium ore during the time of the two World Wars and the Korean War; this mine on the southwestern Kenai Peninsula was the site of the only lode chromite ever mined in Alaska. At Red Mountain, the largest layer of chromite is almost 200-m-long and is as wide as 1.5 m. Platinum group element anomalies are associated with the magmatic chromite in the cumulate ultramafic rocks. The origin of the chromium-rich layers (Fig. 36) remains controversial. Burns (1985) argued that the ultramafic rocks represent the plutonic core.

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Figure 34. A) Orogenic gold deposits of the Willow Creek district in the Peninsular terrane of south-central Alaska are hosted mainly in the southern margin of the slightly older Talkeetna Mountains batholith. B) Workings at the Independence mine were some of the most extensive in the district, mining high-grade gold from a large crack-seal-type quartz vein.
Figure 35. (A) Location of Fe-Cu-Zn VMS deposits in the subduction-accretionary complex rocks, Prince William Sound southern Alaska (after Koski et al., 2007). The deposits are in mainly sedimentary rocks, but adjacent to early Tertiary mafic volcanic units associated with a near-shore spreading ridge, including (B) pillow basalts and (C) feeder dikes.
of the Jurassic Talkeetna Oceanic Arc, exposed along the terrane suture during Cretaceous accretion and uplift. Alternatively, Kusky et al. (2007) suggested that most of the fault-bounded Late Triassic ultramafic massifs may represent bodies formed in an extensional fore-arc supra-subduction zone seaward of the Talkeetna arc. These were then incorporated within the leading edge of the Chugach terrane during its accretion to Wrangellia.

**Southwestern Alaska**

**Epizonal mercury-antimony deposits**

Generally small epizonal mercury and antimony occurrences are scattered throughout rocks of the Kuskokwim basin and, to a lesser extent, surrounding terranes (Sainsbury and MacKevett, 1965; Gray et al., 1997). **Red Devil** (Fig. 37) was the largest of these and the only deposit in Alaska to have produced significant amounts of mercury. The deposits and prospects are mainly present as thin quartz veins in fractures or as breccias in clastic sedimentary rocks, granite sill-dike swarms, and mafic dikes. At Red Devil, however, some of the veins were as much as 1-m-wide and tens of meters in length. Cinnabar and/or stibnite are the dominant sulfides in the veins and breccias, and realgar, orpiment, pyrite, native mercury, and liquid and solid hydrocarbons are also present. Gangue phases include dickite, kaolinite, rare sericite, and carbonate. Anomalous gold is recognized in many of the occurrences (Gray et al., 1990). Dating of hydrothermal sericite indicates metals were deposited ca. 70 Ma,
thus coevally with the late Cretaceous magmatism; there is also a spatial association between many of the occurrences and the granite dikes. Nevertheless, Goldfarb et al. (1990) argued Hg- and Sb-depositing fluids and sulfur were sourced from fluids generated during devolatilization at depth within the Kuskokwim basin.

Gold-bearing deposits

Pluton-hosted gold deposits, the sources for much of the historic placer gold production in southwestern Alaska, are spatially and genetically associated with the latest Cretaceous magmatism throughout the region (Fig. 23, Fig. 37; Bundtzen and Miller, 1997). These deposits have many features that resemble porphyry copper systems, including stockwork style gold and copper mineralization in the cupolas of shallow intrusions. They differ from such deposits, however, by the ilmenite-series chemistry of the host intrusions and lack of well developed porphyry style mineralization, and therefore, at least to some degree, resemble reduced intrusion-related gold systems (Graham et al., 2013). The largest examples of these ca. 70 Ma pluton-hosted deposits are the stockworks and brecciation hosted by syenite, monzonite, and quartz monzonite at Chicken Mountain, and the shear-hosted veins in monzodiorite at Golden Horn (Fig. 38; Bundtzen et al., 1992; Bundtzen and Miller, 1997). The Cu-Au skarns at the Nixon Fork deposit in Ordovician limestones bordering the northern side of the Kuskokwim basin (Fig. 37; Newberry et al., 1997a) and auriferous greisens at the Shotgun prospect at the southern end of the basin (Rombach and Newberry, 2001) both may be additional gold deposits closely related to the same magmatism. The approximately 1000 t Cu and 5 t Au recovered from Nixon Fork is the most significant lode gold and copper production from southwestern Alaska.

A number of small gold-bearing deposits (e.g., Yankee-Gaines Creek, Stuyakok), are hosted within the granite porphyry dike and sill systems, similar to many of the mercury and antimony lodes. They consist of finely disseminated Au-bearing arsenopyrite and arsenate minerals and quartz ± carbonate-sulfide veins (Bundtzen and Miller, 1997). In contrast to the pluton-hosted gold deposits described above, these granite porphyry-related deposits lack appreciable...
Cu and are more strongly associated with linear faults or shear zones (Bundtzen and Miller, 1997).

Historically these gold-bearing deposits did not represent economically significant gold resources. However, the Donlin Creek deposit (Large) was defined as a giant resource in the late 1990s and is the one significant deposit hosted by the sill and dike complexes (Fig. 39A). It is now estimated to contain a world-class resource (measured+indicated+inferred) of 1100 t Au at grades of slightly more than 2 g/t (Nova Gold Resources [novagold.com], September 15, 2015). The giant deposit is located along a dilational zone in a subsidiary fault of the Iditarod-Nixon fault system and at the intersection of this NE-striking secondary structure with an east-west-trending regional anticline. The approximately 3 km x 1.5 km group of orebodies are hosted within an 8 km x 3 km, NNE-trending, discordant and porphyritic rhyolitic to rhyodacitic dike-sill complex that intrudes the Cretaceous Kuskokwim Group sedimentary strata. Radiogenic isotope data presented by Goldfarb et al. (2004) suggested that the sill-dike complex was generated by melting of fine-grained sedimentary rocks deeper within the Kuskokwim basin. Nova Gold Resources (Nova Gold Resources website, September 15, 2015) indicates that the mineralized intrusive rocks continue over a vertical extent of a least 945 m.

Figure 39. The 1100 t Donlin Creek deposit located in the Kuskokwim basin of southwestern Alaska. A) Plan map showing the NE-trending dike swarm, mapped and inferred faults, and surface projections of ore zones, and a cross section with outlines of mineralized intercepts largely hosted within the brittle granite porphyry dikes. Modified from Graham et al. (2013). B) Early, thin pyrite veinlet cutting rhyodacite porphyry. C) Quartz vein with gold-bearing arsenopyrite selvage cutting earlier pyrite vein. D) Late vein with realgar, native arsenic, and stibnite that overprints earlier auriferous arsenopyrite event. Photos courtesy of Scott Petsel from unpublished report of Piekenbrock and Petsel (2003).
Gold- and sulfide-bearing quartz-carbonate veins and veinlets fill north- to northeast-trending extensional fractures in the igneous rocks (Fig. 39B-D) and in the more competent coarse-grained adjacent sedimentary beds (Ebert et al., 2000; Miller et al., 2000). Most of the brittle veins are only a few centimeters in width and the longest veins are about 7 m. The quartz and carbonate veins generally contain less than 3 % arsenopyrite and stibnite, with traces of cinnabar. Gold is refractory within fine-grained, typically needle-like arsenopyrite in the veins (Ebert et al., 2000). Pyrite is generally present as pre-ore stage veinlets. Late fractures can be dominantly realgar, orpiment, and locally, native arsenic. Fine-grained muscovite and illite are the most widespread alteration phases in the dikes and sills, with also carbonization, sulfidation, and kaolinization. The ages of dike rocks ranged from 74.4 ± 0.8 to 70.3 ± 0.2 Ma (40Ar/39Ar; biotite; Szumigala et al., 2000). Mineralization ages between 73.6 ± 0.6 and 67.8 ± 0.3 Ma (40Ar/39Ar; Gray et al., 1997; Szumigala et al., 2000) overlap and extend to slightly younger ages than those of the host igneous rocks.

The genesis of the Donlin Creek deposit is controversial. Ebert et al. (2000) identified features characteristic of a low-sulfidation epithermal deposit, whereas a reduced intrusion-related gold system was favored by Lang and Baker (2001). Goldfarb et al. (2004) suggested the very heavy oxygen isotopes, very light sulfur isotopes, and mixed H2O-CO2 nature of the ore-forming fluids were more consistent with an epizonal orogenic gold deposit. Although the resource at Donlin Creek makes the deposit a true giant and the largest recognized gold-only type deposit in the North American Cordillera, the refractory nature of the gold ore and the lack of infrastructure in the region surrounding the deposit have hindered development of a proposed open-pit mine.

The western Alaska Range has been the focus of much of the gold exploration in Alaska during the past decade. Gold-rich occurrences that have generated significant interest include the Whistler (Large) Au-Cu porphyry (Fig. 40A-B), the Terra low-sulfidation epithermal(?), and the Estelle ilmenite series pluton-hosted polymetallic vein deposits (Fig. 40C, Graham et al., 2013). The deposits formed between about 76 and 68 Ma in association with the younger
period of magmatism within the Kahiltna flysch. The oxidized host diorite at Whistler is associated with the intersection of NE- and NW-striking fault systems, whereas the Estelle pluton and the diorite dike hosting much of the Terra mineralization both are characterized by NNW-oriented vein systems. A paleo-reconstruction of southern Alaska by Graham et al. (2013) at 70 Ma indicates these magmatic-hydrothermal gold systems formed a belt of Late Cretaceous deposits on the more landward side of the Kahiltna flysch basin, relative to the mid-Cretaceous porphyry deposits.

**Platinum/palladium deposits**

Marine sedimentary rocks and volcanic rocks of the small Goodnews terrane were accreted to southwestern Alaska beginning in Early Jurassic (Box, 1985). As an arc-trench system developed, ultramafic rocks of the early Middle Jurassic Goodnews Bay Complex were emplaced into the terrane and they are now exposed about 150 km south of the mouth of the Kuskokwim River. Placer platinum downstream from the intrusive complex (Fig. 37, Fig. 41), recognized for 90 years, represents the largest PGE resource in the United States (Foley et al., 1997). The complex is a concentrically zoned Urals-Alaska type complex with a dunite-wehrlite core, zoned to a narrow clinopyroxenite zone, outward to a hornblende-clinopyroxenite zone, and then into hornfels country rocks (Foley et al., 1997). The dunite is highly serpentinized, with 10-60 % of exposed rock showing olivine altered to serpentine. Platinum is associated with chromite and magnetite in the dunite core, whereas palladium is associated with pyrrhotite and chalcopyrite in the pyroxenite (Mertie, 1976; Southworth and Foley, 1986). Despite the huge placer endowment, no economic PGE lode sources have been recognized; this may reflect erosion of the most significant PGE occurrences or the highly disseminated nature of the PGEs (Mertie, 1976).

In the Nushagak lowlands, in southernmost southwestern Alaska (just south of 60°), an aeromagnetic anomaly identified in 1959, in a region with 300-m-thick Quaternary cover, was subsequently drilled to identify a large mafic-ultramafic intrusion emplaced into sedimentary rocks of uncertain affinity. The composite clinopyroxenite and gabbro bodies contain zones of high-grade titaniferous magnetite, Cu-Zn sulfides, and anomalous platinum (Foley et al., 1997). This Kemuk prospect, dated at ca. 86 Ma (Iriondo et al., 2003), was classified by Foley et al., (1997) as a Urals-Alaska type complex of zoned mafic-ultramafic rocks.

**Alaska to the south of 60 degrees**

The southernmost parts of Alaska, to the south of 60 degrees, contain additional significant metallic mineral resources. The southeastern Alaska panhandle region, due to its complex geology of numerous narrow, lengthy, fault-bounded terranes, is probably the most metal-rich part of Alaska. It includes many pre-, syn-, and post-ac-
cretionary ore deposits formed at different crustal depths over more than 500 m.y. The southern part of southwestern Alaska hosts the giant Pebble porphyry Cu-Au-Mo deposit, which contains the world’s largest known igneous rock-hosted gold resource. Other porphyry and epithermal prospects exists further south along the Alaska Peninsula-Aleutian Island chain.

Southeastern panhandle

The two largest gold producing areas of the 20th century in Alaska were the Juneau gold belt and the Chichagof district. The orogenic gold deposits in the Juneau area occur along a 160-km-long by 5- to 8-km-wide structurally deformed zone centered around the suture between a Jurassic-Cretaceous flysch basin sequence and older parautochthonous terranes, about 15-20 km seaward of the generally slightly older, subduction-related Coast batholith (Fig. 42A-B) Goldfarb et al., 1988). The largest deposits, including those at the historic Alaska-Juneau (Fig. 42C) and Treadwell mines, which yielded slightly less than 200 t Au combined prior to World War II, and at the presently active Kensington mine, are hosted by pre-ore competent intrusive bodies that were emplaced within the metasedimentary rock sequences. The deposits mainly formed along giant shear and tensional vein systems at ca. 56-53 Ma during seismic events associated with changes in Pacific basin plate motions (Goldfarb et al., 1991; Miller et al., 1994). Seaward of the Juneau gold belt, the Chichagof (Fig. 33B) and Hirst-Chichagof orogenic gold deposits were mined in the first half of the 20th century from local shear zones cutting metasedimentary rock sequences. The Juneau Gold Belt of southeastern Alaska was the largest historic gold producer in Alaska. A) The 160-km-long Juneau gold belt, SE Alaska, includes numerous deposits along the length of a complex structural zone, sometimes termed the Coast Range Megalineament (CRM). This zone includes two closely-spaced thrust faults that represent mid-Cretaceous sutures of terranes being added to the Cordilleran orogen. B) Geology of the central part of the Juneau gold belt in the area of the ca. 55 Ma Alaska-Juneau (AJ) and Treadwell deposits. A syndeformational belt of tonalite sills was emplaced 5-10 km landward of the gold deposits between 72 and 58 Ma. Subsequently, post-kinematic plutons of the Coast batholith were emplaced east of the sills. C) Gold ore in AJ pit preferentially hosted in veins in competent diorite bodies. Figure 41. Platinum mine dredge downstream from ultramafic rocks, Salmon River, Goodnews Bay region, southwestern Alaska. Photo courtesy of Gary T. Mason.
mentary rocks of the Chugach accretionary complex. These ca. 50 Ma crack-seal vein systems have been suggested to be products of the same ridge subduction event (Fig. 33F) that led to formation of smaller gold deposits described above in the Chugach-Kenai Mountains (e.g., Haussler et al., 1995).

Allochthonous rocks of the Wrangellia composite terrane in southeastern Alaska host a number of belts of different aged sea-floor VMS deposits that formed far from the present-day Alaskan continental margin. Small metamorphosed and deformed latest Proterozoic and Cambrian and Ordovician to Early Silurian deposits are located in southernmost southeastern Alaska and were mined mainly for their copper, zinc, and precious metals in the early 1900s (Fig. 43; Newberry et al., 1997b). The older occurrences, such as Niblack and Khayyam, represent the oldest recognized mineral deposits in Alaska. A more significant belt of VMS deposits, associated with Late Triassic rift-related bimodal volcanism, stretches for 400 km along much of the panhandle (Fig. 43), and to the north into British Columbia, Canada, where the giant Windy Craggy Cu-Co-Au deposit (190 Mt of 1.6 % Cu and 0.2 g/t Au) is located (Peter and Scott, 1999), although in a park area closed to mine development. The VMS deposits include Castle Island, which was the site from 1963 to 1980 of the only barite mined in Alaska, and of the presently active Greens Creek mine. The latter Zn-Pb-Ag-Au deposit, the United States’ largest silver producer, formed exhalative to replacement style ores in a shallow propagating intra-arc setting along the contact between argillite and mafic flows (Taylor and Johnson, 2010).

Figure 43. Generalized map of terrane boundaries (bold lines) and major VMS deposits, southeastern Alaska. The Late Triassic deposits, including Greens Creek, occur along the length of the SE Alaska panhandle within the Alexander terrane of the Wrangellia composite terrane and are the most significant of the deposits. Modified from Taylor and Johnson (2010).
The southeastern Alaskan panhandle also hosts a number of other important magmatic and magmatic-hydrothermal ores. The Quartz Hill porphyry molybdenum deposit mentioned earlier is hosted by a rift-generated Oligocene felsic stock emplaced into a calc-alkaline subduction-related arc (Ashleman et al., 1997). The deposit, the largest arc-related type porphyry molybdenum deposit in North America, also possesses some features that are more characteristic of the extensional alkali-feldspar rhyolite-granite porphyry molybdenum deposit group and may be more of a hybrid deposit between the two deposit types (e.g., Taylor et al., 2012). This part of Alaska is also one of the type localities for ores that are typically referred to as Urals-Alaska type zoned mafic-ultramafic complexes. The undeveloped, 40-km-wide, roughly syn-accretionary, mid-Cretaceous Klukwan-Duke belt of these intrusions stretches for >500 km along the length of southeastern Alaska (Fig. 5), with 15-20 % Fe in the outer hornblende-pyroxenite part of the complexes and Pt-bearing chromite in the dunite cores (Foley et al., 1997). A group of poorly studied ca. 40-30 Ma composite tonalite, gabbro, and norite bodies, which intrude the Chugach accretionary prism in the northern part of the panhandle, host Cu-Ni-PGE mineralization in the more mafic intrusions. These include the Brady Glacier deposit, closed to mining in a remote part of Glacier Bay National Park, which represents one of the United States’ largest nickel resources (Foley et al., 1997).

Southern Southwestern Alaska and Alaska Peninsula-Aleutian Island Chain

The giant Pebble Cu-Au-Mo deposit, located 320 km southwest of Anchorage and containing the largest gold endowment of any porphyry deposit in the world (3033 t Au grading 0.35 g/t), is associated with ca. 90 Ma intrusive bodies of the Kaskanak batholith emplaced into the Jurassic-Cretaceous Kahlitna flysch (Fig. 44, Lang et al., 2013). It may have been the northernmost of a series of porphyry deposits formed along the landward margin of the Wrangellia composite terrane in the mid-Cretaceous many hundreds of kilometers south of their present latitude (Fig. 8, 31); other deposits of the belt, described above and shown on figure 5, are exposed in the high elevations of the eastern Alaska Range (Goldfarb et al., 2013). The Pebble deposit formed during 10 million years of magmatism, beginning with emplacement of granodiorite and diorite sills, early alkalic intrusions and related breccias, and finally intrusion of the 90 Ma subalkalic granodiorite Kaskanak batholith, with those rocks along the batholith margin hosting the mineralization (Lang et al., 2013). Lang et al. (2013) suggest the large size of the deposit, as well as its high-grade hypogene ore, reflect multiple episodes of magmatic-hydrothermal events, an effective synhydrothermal fault zone for fluid focusing, and overlying hornfels zones in the flysch forming an aquitard to the upward fluid movement.

The Pebble deposit is divided into the Pebble East Zone and Pebble West Zone, which define two associated hydrothermal centers with the east zone dropped 600-900 m in a graben (Kelley et al., 2013). Pebble West extends from the near surface to 500 m depth, whereas Pebble East, below 300-600 m of Late Cretaceous to Eocene sedimentary and volcanic rock cover, continues to depths below 1700 m (Lang et al., 2013). There is a small zone of supergene mineralization above the West Zone orebody, but generally all ore is hypogene. The chalcopyrite-bornite, pyrite, free gold, and electrum are associated with potassic and sodic potassic alteration, with a kaolinite and illite alteration event redistributing the metals. Areas of extremely high-grade Cu and Au mineralization are associated with a zone of relatively late advanced argillic alteration that was focused in the Pebble East Zone. Rhenium-rich molybdenite was introduced during a late quartz vein episode (Lang et al., 2013). Interpretation of reduced-to-pole aeromagnetic data suggests that similar porphyry-style mineralization may exist below cover elsewhere in the northern part of the Alaska Peninsula (Anderson et al., 2013).

To the southwest, the continental arc transitions into the Cenozoic Aleutian oceanic arc. Porphyry and epithermal prospects are widespread along the arcs (e.g., Wilson and Cox, 1983), but the remoteness, lack of infrastructure, and presence of protected wildlife refuges of the region have hindered exploration programs. Low sulfidation epithermal Au-Ag deposits of the Shumagin Islands, such as the Alaska-Apollo deposit, have seen some minor development and production (Wilson et al., 1996).
Figure 44. A) The location of the giant Pebble Cu-Au-Mo porphyry deposit, in largely concealed terrain, showing the surface expression of the Pebble West Zone (red outline) and Pebble East Zone (black outline) in southern south-western Alaska (from Gregory et al., 2013). B) Geologic map of the ca. 90 Ma Pebble porphyry deposit. Courtesy of Karen Kelley, USGS.
Alaska’s relatively short history of mineral exploitation, barely more than 100 years, and the reconnaissance nature of geologic knowledge for much of the State allow the possibility of major undiscovered ore deposits. The known very large ore deposits have been discovered relatively recently (since ~1970) and in the case of the Fort Knox gold deposit, represent a new, previously unexpected mineral deposit type. It is likely that other major metallic mineral resources remain to be discovered in this region of extremely complex and prospective geology. The known large deposits north of 60 degrees are base and precious metal deposits, as are the vast majority of known mines and prospects. For much of modern history, gold has been an important driver of mineral exploration in Alaska, yet a number of the very large deposits, for example Red Dog and Bornite, are base metal deposits containing little or no gold.

The major known metallogenic provinces, for example those hosting the polymetallic base metal deposits of the Brooks Range and eastern Alaska, and the more widespread gold and porphyry deposits spread across different parts of Alaska, reflect ores formed in quite different temporal and tectonic environments. The VMS and clastic-dominated Pb-Zn provinces reflect largely stratabound Paleozoic mineralization formed in ocean basins that is overprinted in many cases by mid-Cretaceous metamorphism. The gold provinces throughout much of Alaska and porphyry belts in the southern part of the State primarily reflect mid-Cretaceous to Eocene tectonism along active continental margins. Both metamorphism and magmatism may have been significant in the formation of various lode gold deposits; erosion of lodes has led to many large and productive placer gold fields.

Major brownfield and greenfield discoveries are both likely to be part of Alaska’s exploration future. Brownfield developments will be highly influenced by socioeconomic issues. The Fort Knox deposit, for example, was explored and developed at the site of a gold occurrence known for almost 100 years, but a favorable infrastructure near the town of Fairbanks and a local population that mainly supported the nearby mining activity were critical for success. Giant mineral deposits are now recognized at Donlin Creek, Money Knob, and Pebble, but issues of infrastructure, metal price, and/or potential environmental effects are impacting their additional exploration and potential development. The successful model of sustainable resource development in the Red Dog district, with beneficial inclusion of the Native Alaskans in all stages of activity, provides an example that could be followed for future mining of other large tonnage deposits in many parts of the State. Pebble and Pogo represent recent greenfield exploration successes that indicate numerous giant deposits still remain to be discovered throughout Alaska, particularly in the relatively poorly understood areas of extensive young cover. State-of-the-art approaches in exploration geochemistry, remote sensing, and particularly geophysical methods will be required for better defining geology and structure in many of these areas of cover and identifying the most favorable areas for discovery of important resources.
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Deposits

- Diamonds

Size and activity

- Active mine
- Important deposit
- Very large
- Large
- Potentially large

Types of deposits:

- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Locations:

- Greenland
- Alaska
CHAPTER 2

MINERAL DEPOSITS OF ARCTIC CANADA

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Natural Resources Canada. Geological Survey of Canada (GSC)
INTRODUCTION

Geological Setting

This introduction gives a brief overview of the geological setting for the mineral deposits of Arctic Canada (Figure 1). Arctic Canada can be divided into:

- The cratonic roots of the Canadian shield which contains much of the gold, copper, nickel, iron, uranium, rare earth elements and diamonds.

- The bounding Mesoproterozoic to Phanerozoic platform, basins and accreted terrains of the Canadian Cordillera which extend into the High Arctic and contain much of the zinc, lead, gold, silver, copper, molybdenum and tungsten.

Deposits are characterized as small, medium, large, very large and potentially large according to the method described by Eilu et al. (2007) and summarized in the introduction to this volume.

The Canadian shield encompasses four Archean cratons:

- The Slave craton is the oldest: it lies to the north-west and is bound by Paleoproterozoic orogens: the Thelon-Taltson orogen to the east and the Wopmay orogen to the west. The Slave craton is divisible into Hadean to Mesoarchean (4.0 to 2.8 Ga) crust primarily located in the west, and Neoarchean crust and supracrustal rocks which dominate the rest of the craton and also unconformably overlie the Mesoarchean and older basement. The Neoarchean is divided into supracrustal rocks consisting of greenstone belts, collectively referred to as the Yellowknife Supergroup, and overlying turbidites. These are succeeded by felsic to intermediate intrusive suites and by mafic dyke swarms that range from 2.2 to 0.73 Ga. The Slave craton is associated with orogenic gold (12 deposits/camps), volcanogenic massive sulphides (VMS: three deposits, all large, Hackett River), five diamond-rich kimberlites (notably the early producers Ekati and Diavik) and a large rare earth mineral deposit (Nechalacho).

- The Rae and Hearne cratons lie to the east, underlying most of the remaining parts of the Canadian shield across the Canadian Arctic. These contain Mesoarchean and older roots, and cover and plutonic suites of Neoarchean age, all overprinted by the Paleoproterozoic Trans-Hudson Orogeny. Supracrustal rocks are significant in the Neoarchean Rae craton: they contain important resources of iron (Mary River and Roche Bay), orogenic gold of Paleoproterozoic age (3), uranium, associated with a sub-Paleoproterozoic unconformity (four, including the large Kiggavik-Andrew Lake trend), nickel (1) and commercially significant kimberlites (2). Deposit types in the Hearne craton include Ni-Cu-PGE (2), uranium (1), VMS (1) and one large orogenic gold deposit (Meliadine).

- The Superior craton is mostly located S of 60°N but is also exposed in the northern extremity of Quebec. It is bound to the north by the Cape Smith Belt, part of the circum-Superior Trans-Hudson orogen. This belt is noted for its mafic and ultramafic rocks that contain important resources of nickel, copper and platinum group minerals (3 large deposits: Raglan, West Raglan, and Nunavik). The other significant Paleoproterozoic belt is represented by the Wopmay orogen which lies west of the Slave Craton. This features an eastern sedimentary belt and, to the west, the Great Bear Batholith. Important resources in the latter include iron oxide-copper-gold (2 deposits), polymetallic veins (2) and vein uranium (1).

- The Precambrian cratons and their cover of Paleoproterozoic basins are fringed to the north and west by widespread shelf carbonate deposition that begins in the Mesoproterozoic and
continues through the upper Paleozoic. These rocks host Mississippi Valley-type Zn-Pb deposits including large past producers near Great Slave Lake (Pine Point), on northern Baffin Island (Nanisivik) and in the Arctic Islands (Polaris). As yet undeveloped resources are located in the Mackenzie Mountains of the western Northwest Territories (four, of which Gayna River and Prairie Creek are large). Also present in this realm are two iron deposits including the very large Crest deposit in Neoproterozoic strata.

Southwestwards the shelf carbonate succession gives way to Cambrian to Devonian deep water sediments of the Selwyn Basin including shale, chert, carbonate and turbidites. The line of facies change lies in the western Northwest Territories and eastern Yukon. Important resources are represented by sedimentary-exhalative (SEDEX) Zn-Pb mineralization of which there are five deposits in the Yukon (Anvil, Tom and Jason, which are large and Howard’s Pass, which is potentially very large). There are also four VMS camps in the Selwyn Basin, including the large Hasselberg deposit and the large Wolverine deposit in the Finlayson camp.

The western part of the Yukon is dominated by Jurassic and Cretaceous accreted terranes and by associated felsic to intermediate intrusive rocks. This is a key realm for gold (14 deposits), polymetallic Ag-Pb-Zn veins (5, notably Keno Hill, large) and Ni-Cu-PGE (1). The gold deposits are classified into epithermal (4, including the large Golden Revenue deposit), intrusion-related (3), orogenic (2, including Coffee (large), manto-type (2) and iron-oxide-copper-gold (3, one of them very large, Minto). There are also deposits associated with Mesozoic intrusives, tungsten and copper skarns (6 deposits, one large: Mactung), and Cu-Mo porphyry (5, two of them large: Logtung and Red Mountain). Finally, among the many resources of the Yukon, there are eleven gold placer districts of which the Klondike is most significant.

For the purposes of representing ore deposits in the Canadian Arctic on a 1:5 million scale map, selected spatially distinct mineralized zones have commonly been grouped as one deposit (Figure 1). In this context, a deposit is defined as a cluster of genetically related occurrences on a property being investigated by one mineral industry organization (owners, operators, partners) in anticipation of these being mined at a profit as part of a single mine plan. Production and resource figures for these zones have been combined in the database, with a citation given for the source of the resource figures. Added to these deposits, many of which are actively being developed, there are many small, short-lived past producers which are historically well-known but which would not have been mined under the current regulatory environment.

History of Mineral Exploration

The history of mineral exploration in Arctic Canada has its beginnings in the search for placer gold in the Yukon. Although prospecting had uncovered gold along the Yukon River as early as 1883, a report by George Dawson identified the unglaciated areas of west central Yukon as having the greatest potential. Significant gold was discovered in river gravels of Bonanza Creek in August 1896 by George and Kate Carmack, Skookum Jim and Dawson Charlie. This became widely known and by July 1897 there was a major gold rush into the Klondike from the west coast of the USA and from many parts of Canada. In total 30,000 to 40,000 would-be miners entered the region from 1897 to 1899 (Berton, 2001). The rush came to an end when new gold discoveries were reported from the Atlin area of northern British Columbia (in 1898) and from Nome, Alaska (in 1899).

Prospecting activity uncovered new hard-rock gold deposits. In the 1890s this included reconnaissance expeditions by staff of the Geological Survey of Canada. These expeditions and others by prospectors located copper at Carmacks, Yukon by George Dawson (1887) (Kent et al., 2014), polymetallic silver-lead-zinc at Quartz Lake, Yukon (1892), intrusion-related gold at Dublin Gulch (1895) and Lone Star (1897), and silver-lead-zinc at Keno Hill (1901). Further afield, new indications of nickel mineralization were identified by Albert Low in northern Quebec (1898), gold along the Yellowknife River on the north side of Great Slave Lake (1898), zinc and lead sampled by Robert Bell at Pine Point south of Great Slave Lake (1899), and copper, uranium and cobalt at Echo Bay on the east side of Great Bear Lake first noticed by James MacIntosh Bell (1900). The Mississippi Valley-type deposits at Pine Point were low in silver and thus of limited interest to the mining community at that time. It would be sixty-six years before the
Figure 1. Geological location map featuring named deposits of large and very large size across Arctic Canada.

Table 1: Overview of the very large and large deposits in Canada north of 60°

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Status</th>
<th>Size</th>
<th>Genetic type</th>
<th>Main metals</th>
<th>Total tonnage - Mt (Mixed)</th>
<th>Grades</th>
</tr>
</thead>
<tbody>
<tr>
<td>Athabasca</td>
<td>Not exploited</td>
<td>Large</td>
<td>Unconformity</td>
<td>Cu, Mo, Au, Ag</td>
<td>7.07</td>
<td>0.23% Cu, 0.03% Mo</td>
</tr>
<tr>
<td>Canolino</td>
<td>Not exploited</td>
<td>Very large</td>
<td>Porphyry (Cu, Au), W, Sn, Ag</td>
<td>Cu, Mo, Au, Ag</td>
<td>27.026</td>
<td>0.16% Cu, 0.03% Mo</td>
</tr>
<tr>
<td>Coffeey</td>
<td>Not exploited</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>92.98</td>
<td>1.4 ppm Au</td>
</tr>
<tr>
<td>Con Mine</td>
<td>Closed mine</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>13.71 (13.7)</td>
<td>17.5 ppm Au</td>
</tr>
<tr>
<td>Courtena Lake</td>
<td>Renewed exploration</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>156.446 (6.17)</td>
<td>2.3 ppm Au</td>
</tr>
<tr>
<td>Crew</td>
<td>Not exploited</td>
<td>Very large</td>
<td>Stromatolite</td>
<td>Fe</td>
<td>3200</td>
<td>43.0% Fe</td>
</tr>
<tr>
<td>Fasano</td>
<td>Closed mine</td>
<td>Large</td>
<td>Sedimentary chertnict</td>
<td>Zn, Pb, Ag (Au)</td>
<td>56.0 (12.26)</td>
<td>6.5% Zn, 2% Pb, 3% Sn, 35 ppm Ag, 0.3 ppm Au</td>
</tr>
<tr>
<td>Fogo Lake</td>
<td>Not exploited</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu (Cu, Pt)</td>
<td>46</td>
<td>0.7% Ni, 1% Cu, 0.06% Co, 1.3 ppm Pt, 0.2 ppm Pt</td>
</tr>
<tr>
<td>Goose Mine</td>
<td>Not exploited</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>24.76</td>
<td>4.6 ppm Au</td>
</tr>
<tr>
<td>Hackett River</td>
<td>Not exploited</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Pb, Cu, Ag (Au)</td>
<td>87</td>
<td>3.8% Zn, 0.5% Pb, 0.4% Cu, 144 ppm Ag, 0.23 ppm Au</td>
</tr>
<tr>
<td>Hinesberg</td>
<td>Not exploited</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Pb, Ag</td>
<td>4.1</td>
<td>0.2% Zn, 1.8% Pb, 84 ppm Au</td>
</tr>
<tr>
<td>High Lake</td>
<td>Not exploited</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Cu, Pb (Ag, Au)</td>
<td>14</td>
<td>3.5% Zn, 0.4% Pb, 2.8% Cu, 86 ppm Ag, 0.2 ppm Au</td>
</tr>
<tr>
<td>Howards Pass</td>
<td>Not exploited</td>
<td>Large</td>
<td>Sedimentary chertnict</td>
<td>Zn, Pb</td>
<td>366.5</td>
<td>4.9% Zn, 1.6% Pb</td>
</tr>
<tr>
<td>Inuk Lake</td>
<td>Not exploited</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Cu, Pb, (Ag, Au)</td>
<td>14.6</td>
<td>12.1% Zn, 1.4% Pb, 2.3% Cu, 73 ppm Ag, 0.2 ppm Au</td>
</tr>
<tr>
<td>Kenning Silvert</td>
<td>Active mine</td>
<td>Large</td>
<td>AlPb-Zn veins</td>
<td>Ag, Pb, Zn</td>
<td>7.214 (6.947)</td>
<td>4.4% Zn, 0.3% Pb, 1107 ppm Ag</td>
</tr>
<tr>
<td>Kudz Ze-Khayy</td>
<td>Not exploited</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Pb, Cu, Ag (Au)</td>
<td>14.55</td>
<td>5.9% Zn, 0.5% Pb, 0.9% Cu, 121 ppm Ag, 1.3 ppm Au</td>
</tr>
<tr>
<td>Losinggi</td>
<td>Not exploited</td>
<td>Large</td>
<td>Porphyry (Cu, Au, Mo, W, Sn, Ag)</td>
<td>Ni, Mo</td>
<td>424.6</td>
<td>0.02% W, 0.02% Mo</td>
</tr>
<tr>
<td>Lupin Mine</td>
<td>Closed mine</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>12.8 (11.1)</td>
<td>10 ppm Au</td>
</tr>
<tr>
<td>Macung</td>
<td>Not exploited</td>
<td>Large</td>
<td>Blast (Zn-Pb-Ag, Cu, Au, Fe, W)</td>
<td>W</td>
<td>44.886</td>
<td>0.73% W</td>
</tr>
<tr>
<td>Mary River 1</td>
<td>Active mine</td>
<td>Large</td>
<td>Algoma-style iron formation</td>
<td>Fe</td>
<td>631</td>
<td>66.5% Fe</td>
</tr>
<tr>
<td>Mary River 2 &amp; 3</td>
<td>Not exploited</td>
<td>Large</td>
<td>Algoma-style iron formation</td>
<td>Fe</td>
<td>362</td>
<td>65.9% Fe</td>
</tr>
<tr>
<td>Meadjavale Mine</td>
<td>Active mine</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>27.407</td>
<td>3.3 ppm Au</td>
</tr>
<tr>
<td>Melodie</td>
<td>Not exploited</td>
<td>Large</td>
<td>Orogenic gold</td>
<td>Au</td>
<td>43.273</td>
<td>6.5 ppm Au</td>
</tr>
<tr>
<td>Mine</td>
<td>Active mine</td>
<td>Large</td>
<td>IOCG to porphyry</td>
<td>Cu (Au, Ag)</td>
<td>110.144 (53.72)</td>
<td>1.7% Cu, 4.9 ppm Ag, 0.6 ppm Au</td>
</tr>
<tr>
<td>Nacterda Mine</td>
<td>Closed mine</td>
<td>Large</td>
<td>Mt-K toish type Ch (Zn, Pb)</td>
<td>Zn, Pb (Ag)</td>
<td>17.352 (17.54)</td>
<td>9.0% Zn, 0.7% Pb, 41 ppm Ag</td>
</tr>
<tr>
<td>Nacthaloic</td>
<td>Not exploited</td>
<td>Large</td>
<td>Pentakelite rock associated rare metals</td>
<td>REE</td>
<td>304.63</td>
<td>2213 ppm Nb, 1.18% Y, 146 ppm Ta, 1.81% Zr</td>
</tr>
<tr>
<td>Nickel King</td>
<td>Not exploited</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu (Cu)</td>
<td>44.172</td>
<td>0.4% Ni, 0.08 Cu, 2.02% Co</td>
</tr>
<tr>
<td>Nauyack Mine</td>
<td>Active mine</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu, (Cu, Pt)</td>
<td>27.146</td>
<td>0.9% Ni, 1.1% Cu, 0.08% Co, 2.2 ppm Pt, 0.5 ppm Pt</td>
</tr>
<tr>
<td>Pine Point</td>
<td>Closed mine</td>
<td>Large</td>
<td>Mt-K toish type (Zn, Pb)</td>
<td>Zn, Pb</td>
<td>100.88 (64.28)</td>
<td>5.8% Zn, 2.5% Pb</td>
</tr>
<tr>
<td>Polak Mine</td>
<td>Closed mine</td>
<td>Large</td>
<td>Mt-K toish type (Zn, Pb)</td>
<td>Zn, Pb</td>
<td>20.157 (10.157)</td>
<td>13.4% Zn, 3.6% Pb</td>
</tr>
<tr>
<td>Polia Creek</td>
<td>Closed mine</td>
<td>Large</td>
<td>Mt-K toish type (Zn, Pb)</td>
<td>Zn, Pb (Ag)</td>
<td>11.87</td>
<td>12.9% Zn, 10.3% Pb, 0.5% Cu, 197 ppm Ag</td>
</tr>
<tr>
<td>Raglan Mine</td>
<td>Active mine</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu, (Pt)</td>
<td>42.03 (8.89)</td>
<td>3.2% Ni, 0.9% Cu</td>
</tr>
<tr>
<td>Red Mountain</td>
<td>Not exploited</td>
<td>Large</td>
<td>Porphyry (Cu, Au, Mo, W, Sn, Ag)</td>
<td>Mo</td>
<td>167</td>
<td>0.1% Mo</td>
</tr>
<tr>
<td>Roche Bay C</td>
<td>Not exploited</td>
<td>Large</td>
<td>Algoma-style iron formation</td>
<td>Fe</td>
<td>567.3</td>
<td>26.4% Fe, 0.08% P2O5</td>
</tr>
<tr>
<td>Tom and Jackson</td>
<td>Not exploited</td>
<td>Large</td>
<td>Sedimentary chertnict</td>
<td>Zn, Pb, Ag</td>
<td>30.99</td>
<td>0.5% Zn, 3.8% Pb, 39 ppm Ag</td>
</tr>
<tr>
<td>Welshman Mine</td>
<td>Renewed exploration</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu, Pt, Ag (Cu, Au)</td>
<td>431.28 (1.17)</td>
<td>0.3% Ni, 0.3% Cu, 0.02% Co, 0.3 ppm Pt, 0.4 ppm Pt</td>
</tr>
<tr>
<td>West Raglan</td>
<td>Not exploited</td>
<td>Large</td>
<td>Magnetite Ni-Cu-PGE</td>
<td>Ni, Cu, (Pt, Pt)</td>
<td>10</td>
<td>2% Ni</td>
</tr>
<tr>
<td>Wilkowana Mine</td>
<td>Active mine</td>
<td>Large</td>
<td>VMS</td>
<td>Zn, Cu, Pb, Ag (Au)</td>
<td>6.154</td>
<td>12.2% Zn, 1.6% Pb, 1.2% Cu, 303 ppm Ag, 1.7 ppm Au</td>
</tr>
</tbody>
</table>
area would become commercially viable (Siega and Gann, 2014).

The introduction of float-planes from 1925, was a significant boost to the mineral exploration business (Zaslow, 1975). This allowed access to large areas of remote country and was largely responsible for the exploration discoveries of the 1930s. The prospector Gilbert LaBine was the first to investigate the mineral showings on Great Bear Lake: in 1930 this led to the discovery of high grade pitchblende and silver ore at what would become the Eldorado mine (Zaslow, 1975). A small staking rush identified vein uranium at Rayrock, also near Great Bear Lake, in 1934, and of copper-gold mineralization at NICO. Production at the Eldorado Mine began in 1933 followed by the Contact Lake silver mine in 1936. The Eldorado mine was closed in 1940, but was soon reopened to supply uranium to the war effort. It remained mostly active down to 1982.

The discovery of gold near Yellowknife in 1898 was not, at the time, considered significant, partly because of the remote location of the area. However, the success of mine development at Great Bear Lake inspired new prospecting over a wide area of the Northwest Territories. Renewed prospecting took place in the early 1930s, which led to the discovery, in 1935, of the deposits which were developed in the Giant and Con mines (Duke et al, 1991; Goucher, 1991) and to the establishment of the Yellowknife town site from 1936. As many as five mines were operating in the late 1930s. The outbreak of WWII, however, brought much of this activity to an end.

The 1960s and 1970s were notable for the exploration for porphyry copper-molybdenum in the Yukon following on the success of new exploration models in British Columbia (Sinclair, 2007). This led to discoveries at Red Mountain and Casino (both in 1967), which precipitated a staking rush and later the discovery of Cash (1975) and the Logtung tungsten-molybdenum deposit (1976). New properties associated with iron oxide-copper-gold (IOCG) mineralization were discovered in this period, including the Sue Dianne deposit (1974) which has a noteworthy radiometric anomaly, Pagisteel in the Mackenzie Mountains and the Minto deposit (1971), Yukon, which was staked on the basis of a stream-sediment geochemical anomaly (Mercer et al, 2012). Mississippi Valley type deposits also became commercially viable in this period, notably Pine Point (1965), Polaris in the Arctic Islands (1970) which was discovered by drilling a gravity anomaly (Dewing et al., 2007a), Gayna River (1974), and Nanisivik on Baffin Island (1976). Other noteworthy discoveries included the following sedimentary exhalative (sedex) deposits: the Anvil deposits (1953, 1965), Tom and Jason (1951, 1974) and, especially, the large Howard's Pass deposit (1972) which was discovered during follow-up of stream sediment geochemical results. Announcement of the Howard's Pass discovery precipitated a major staking rush across the central Yukon (O'Donnell, 2009). Volcanogenic massive sulphide deposits were also being discovered in the Yukon, e.g. Hasselberg (1953) and Hart River (1955), and notably, on the Canadian shield, Hacket River (1966).

Other developments in this latest period included the discovery of orogenic gold deposits in the Hope Bay belt in the northeastern Slave craton (1992-1995), Hyland gold (1981) which started out as a lead-zinc property (Armitage and Gray, 2008).

The latest significant development in the mineral exploration of Arctic Canada has been the discovery of commercially significant diamond-bearing kimberlites. This was the vision of two men, Charles Fipke and Stewart Blusson, who tracked kimberlite indicator minerals extracted from eskers and in so doing pin-pointed the favourable kimberlite source in bedrock near Lac de Gras in the central Slave craton. This was an endeavor of ten years ending with drilling of the first diamond-bearing kimberlite in 1991 (Carlson et al., 2015). The announcement of their discovery precipitated one of the biggest staking rushes in Canadian history. As well as the original Ekati property, other kimberlite prospects of economic significance were located in several parts of the Arctic, including Snap Lake (1994) and Diavik (1995) in Northwest Territories and Qilalugaq (2000-2005) and Chidliak (2005) in Nunavut.

**NEOARCHEAN IRON**

Two Neoarchean iron deposits, each of large dimensions, are located in the Rae craton of northeastern Nunavut. The Mary River deposit (Fig. 1) is located on northern Baffin Island, 160 km S of Pond Inlet, 300 km N of Sanirajak, and 1000 km NW of Iqaluit (the capital of Nunavut). Access is by fixed wing to an airstrip on the property, or by ski-or float-equipped aircraft to nearby Sheardown Lake. The second major deposit is the Roche Bay iron deposit located 60 km SW of the settlement of Sanirajak and immediately west of Roche Bay on eastern Melville Peninsula (Figure 2).

**Mary River (Large)**

Iron ore of commercial significance was proven by drilling at Mary River as early as 1965. There was no new interest until 2004 when additional drilling was completed. A revised estimate of the resource was made available in 2006. The current owner is Baffinland Iron Mines Corporation. The Mary River group is named for a collection of supracrustal outliers in the northwestern part of Baffin Island. The tectonic setting is one in which the Mary River Group (2.74 to 2.68 Ga) and associated deposits occur in west- and northeast-trending synforms overlying older Mesoproterozoic basement. Jackson (2000) and Johns and Young (2006) have correlated these belts with the Prince Albert and Woodburn groups of Melville Peninsula and with similar rocks in northwest Greenland, a tectonic entity that is identified as the Committee Orogen (Jackson, 1966). However, these relationships are now in some doubt. The Mary River group overlies a basement complex that includes foliated and nebulitic granite and granodiorite (3.0 to 2.8 Ga; Johns and Young, 2006). The contact relationship is thought to be unconformable but in places is also faulted, mylonitic, intrusive and migmatic. The Mary River Group is succeeded regionally by the Paleoproterozoic Piling Group (central Baffin Island), generally unmetamorphosed Mesoproterozoic strata of Borden Basin, dykes of the Mackenzie (1270 Ma) and Franklin (720 Ma) swarms, and flat-lying or tilted Ordovician carbonates.

The thickness of the Mary River Group in the vicinity of the iron deposits is considered to be of the order of 2000 - 4000 m. The metamorphic grade ranges from greenschist to upper amphibolite facies. Jackson (2000) has identified the following five units within the Mary River Group (from the base upwards): 1) metapelitic, mafic metavolcanics; lenses of conglomerate; 2) quartzite; metarhyolite, dacite (2718±5-3 Ma),

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1 The sizes of the deposits described in this chapter are grouped according to the criteria described by Eilu et al. (2007).
mafic metavolcanics; oxide facies iron formation (and ore deposits); 3) conglomerate, breccia; greywacke, metapelite; metavolcanics; pyribo-lite; iron formation; 4) local conglomerate, breccia; greywacke, metapelite; 5) local conglomerate, breccia; metavolcanics with meta-anorthosite, meta-gabbro, pyribo-lite.

Local unconformities have been proposed to underlie the conglomerates within each unit.

Layering of iron formation with other lithofacies is common and has been attributed to a mixture of both depositional facies variation and tectonic interleaving. Varieties of iron formation include oxide, silicate, pelitic and calcitic/ferroan dolomitic carbonate. Oxide and pelitic facies iron formation also contain small amounts of disseminated pyrite and pyrrhotite. The greatest thickness of iron formation occurs in the vicinity of the ore bodies: 52 - 195 m thick and traceable.
for up to 3.8 km (Figure 3). The ore zones are, however, generally lenticular in shape. The ore zone mineralogy is mostly hematite–magnetite grading to hematite. Subsidiary phases include grunerite, anthophyllite, clinochlore, quartz, garnet, pyrite, pyrrhotite, chalcopyrite and covellite (Wahl et al., 2011).

**Roche Bay (Large)**

The Roche Bay deposits are classified as Algoma-type iron formation. Five deposits are located here and labelled as zones A, B, C (Large), D and E. Exploration in the period 2006–2011 focussed on Zone C: its dimensions are 5 km in strike length, 160 m thickness and a dip of 70° to the southwest. The A/B Zone is 1400 - 2000 m long and 120 - 150 m thick.

The oldest rocks of the eastern Melville Peninsula area are granitized paragneisses of the Amitioke Gneiss Complex. These were identified as basement to the iron formation-bearing Prince Albert Group by company geologists (Palmer, 2007), who also recognized an intrusive relationship. The type Prince Albert Group in the Prince Albert Hills of western Melville Peninsula has, more recently, been redefined, and dated to 3.20 to 2.77 Ga (Corrigan et al., 2013), with uncertain correlation to the Archean sediments at Roche Bay. Typical Archean supracrustal rocks near the deposits include metasedimentary and metavolcanic rocks. The metasediments feature iron formation, quartzite (metachert and meta-quartz arenite), biotite schist, argillite and meta-conglomerate (Palmer, 2007, Schau, 1981). The metavolcanics include meta-andesite, meta-rhyodacite, meta-rhyolite, and felsic breccia. Mafic igneous rocks, including metagabbro and “metabasalt” sills of the Tasijuaq Gabbro Complex, intrude the supracrustal rocks.

The metamorphic grade ranges from upper greenschist to lower amphibolite facies. The structure of the area features two phases of folding with the longer wavelength sets being isoclinal. There are two fabric sets: S1, which is parallel to basalt pillows and a younger S2 which is axial planar to regional folds. The iron formation shows a lamellar interlayering of magnetite and quartz (metachert) with individual layers averaging 0.5 - 2 cm thick. Host-rock minerals include magnetite, quartz, amphiboles (grunerite, actinolite, hornblende), micas (biotite, muscovite), chlorite, and sulphides (mostly pyrite and pyrrhotite, minor chalcopyrite and arsenopyrite) (Schau, 1981).

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Figure 3. Mary River: View to the northwest of the No. 1 deposit. Length of the deposit is 2.5 km and consists of 68.2% average iron as hematite and magnetite. GSC 1995-201A.
NEOARCHEAN VOLCANOGENIC MASSIVE SULPHIDE (VMS)

The Slave craton is an important host for ore deposits in Arctic Canada. Lying in the northwestern part of the Canadian shield, it is bound to the east by the Paleoproterozoic Taltson-Thelon orogen and to the west by the Wopmay orogeny. The Slave craton features a basement of Eoarchean to Mesoarchean granitoid rocks and several supracrustal units (4.0 to 2.83 Ga). Unconformable on these, there are supracrustal units collectively referred to as the Yellowknife Supergroup (2.74 - 2.62 Ga). This contains quartzite and ironstone in the lower part, tholeiitic greenstone, arc rocks, and, in the upper part, turbidites. This package was intruded by a “bloom” of granites at 2.95 to 2.58 Ga and was intensely deformed at 2.64 to 2.58 Ga. The mineral deposits of the Slave craton include volcanogenic massive sulphides in the Yellowknife Supergroup, shear-zone hosted gold and iron-formation hosted gold. Also present are diamond-bearing kimberlites (described near the end of this account) of Cambrian to Eocene age (Bleeker and Hall, 2007).

Three massive sulphide deposits are located in the northern part of the Slave craton of western Nunavut (Figure 4). The large Izok Lake deposit in the northwestern Slave is 265 km S of the community of Kugluktuk (Coppermine) on Coronation Gulf. The Hackett River deposit (large) in the Hackett River greenstone belt of the northeastern Slave is 485 km NE of Yellowknife. The large High Lake deposit in the High Lake greenstone belt is located in the northern Slave craton 40 km S of Coronation Gulf. Elsewhere, the Heninga deposit is located in the Rankin-Ennadai greenstone belt of the Hearne craton W of Hudson Bay (Figure 5).

Izok Lake (Large)

High-grade massive sulphide boulders were discovered in the Izok Lake area by Texasgulf Inc. in 1974. Drilling of the adjacent lake soon led to the discovery of four massive sulphide lenses (Northwest, North, Central West, Central East). A fifth lens (Inuksuk) was discovered in 1992.

Two regional map units are present in the area. The lower of these, the Point Lake Formation, consists of marine mafic tholeiite and felsic calc-alkaline volcanics and volcaniclastic rocks. This is overlain by the Contwoyto Formation which consists of iron formation and greywacke turbidites. These units are intruded by granitoid rocks and by Mesoproterozoic Mackenzie dykes.

The Point Lake Formation at Izok Lake features felsic to mafic flows and some calcareous metasediments. The felsic volcanics are mostly porphyritic rhyolite and calc-alkaline volcanics and volcaniclastic rocks. Volcanic rocks in the immediate vicinity of the ore lenses include pyroclastic aphyric rhyolite and quartz-phryic rhyolite, often hydrothermally altered. Other significant rock units in the same area include andesite dykes and flows, volcaniclastic rocks, pillow basalt and gabbro dykes and sills. Iron formation is also present and often contains disseminated sulphides (Morrison 2004). Deformation in the area is primarily represented by a layer-parallel strain fabric with a superimposed crenulation cleavage. Metamorphism is typically high temperature and low pressure based on assemblages that contain sillimanite and cordierite-spinel-corundum. Alteration is most intense around the ore deposits and includes sericitization (muscovite), chloritization (chlorite-biotite-cordierite), biotite and silicification (Morrison 2004).
The ore lenses are mineralogically complex and include combinations of the following sulphides: pyrite, sphalerite, chalcopyrite, pyrrhotite and galena. A copper-rich stockwork zone is also present, beneath the massive sulphide ores, for example under the Central West lens. Sizes range up to 570 x 200 m and 125 m thick in the case of the Central lenses (Morrison 2004). The ore deposits have been interpreted as forming in a sub-sea floor replacement setting rather than as a sea-floor mound. This is drawn from the observation that there are no exhalites associated with the ore horizons. Franklin et al (2005) have identified the Izok Lake deposit
as a bimodal-mafic type volcanogenic massive sulphide.

**Hackett River (Large)**

The Hackett River property was acquired by Glencore Canada Corporation in 2011. It lies within the Neoarchean Hackett River greenstone belt of the northeastern Slave craton, close to its unconformable contact with the Paleoproterozoic Goulburn Group of Kilohigok Basin. A volcanic-dominated unit, the Hackett River Group dominates the lower part of the section. It is correlated with the Banting Group of the Yellowknife area (Bleeker and Hall, 2007) and is subdivided as follows. At the base is the Siorak Formation...
(biotite schist, quartzofeldspathic gneiss, sericitic schist, amphibole gneiss) overlain by the Nauna Formation (volcanic flows, in part pillowed, mafic and felsic pyroclastics). This is succeeded by the Iqnerit Formation, consisting of dacitic and andesitic flows and pyroclastics (Clemmer et al., 2013). Intercalated with the flows are “white smoker” dolomitic and calcitic carbonates (Bleeker and Hall, 2007), iron formation, and sulphidic volcaniclastic rocks. The Iqnerit Formation is the host of the known VMS deposits (Figure 6).

Turbidites of the Beechey Lake Group overlie the sequence described above. Characteristic rocks include greywacke, mudstone and minor carbonaceous mudstone, iron formation, chert and pyroclastic rocks. The Beechey Lake Group is correlated with Burwash Formation turbidites of the Yellowknife area and other widely represented metasediments of the Slave craton (Bleeker and Hall, 2007). Two granitoid suites are present in the Hackett River area. The Mara River complex features granitoid gneiss, migmatite and anatectite derived from the Yellowknife Supergroup. More discrete plutonic bodies are found in the intrusives of the Regan Intrusive suite: granite, granodiorite, tonalite and quartz diorite plutons. These display metamorphic aureoles in the enclosing rocks. Largely post-tectonic, these are probably part of the “granite bloom” of Bleeker and Hall (2007; 2595-2585 Ma).

VMS deposits in the Hackett River belt are spatially associated with synvolcanic intrusions and include the A deposit, East Cleaver, Boot Lake, and Cleaver Low Grade in the north, and the Yava and Musk deposits in the south (Bleeker and Hall, 2007). Alteration associated with the deposits includes 1) sillimanite-biotite-garnet-quartz; 2) sericite-quartz; and 3) anthophyllite- cordierite (Clemmer et al., 2013). Deformation has produced upright and recumbent to overturned folds. There is also late-stage reverse faulting. Cu-Pb-Zn and Ag ternary plots have been employed to infer an arc-like VMS setting (Bleeker and Hall, 2007). These deposits are classified as bimodal-felsic VMS (Franklin et al., 2005).

**High Lake (Large)**

The High Lake greenstone belt, 70 km in strike length and 5 - 25 km wide, is divided into a central metasedimentary belt bound to both east and west by volcanic domains. The western domain is dominated by intermediate and felsic volcanic rocks and volcaniclastic rocks with an age range of 2705±1 Ma to 2695±3 Ma (Henderson et al, 1995). Gossans, massive sulphides and gold occurrences are common. The central sedimentary domain comprises mostly slate, siltstone, greywacke, volcaniclastic rocks and chemical carbonate metasediments. It is the host for the Ulu gold deposit and other epithermal arsenide-gold occurrences. Its age range

Figure 6. Hackett River: massive sulphide with up to 30 % buckshot pyrite in a matrix of dark red sphalerite and tremolite-altered tuff clasts. GSC 2015-111.
is 2616±3 Ma to 2612±3 Ma (Henderson et al, 1995; Henderson et al, 2000). Mafic and intermediate volcanic and volcaniclastic rocks are dominant in the eastern domain. It is geologically and metallogenically similar to the western domain.

Deformation fabrics include a bedding parallel foliation, NNE-striking map scale folds, a ubiquitous, steeply-dipping NNE-striking planar fabric (that overprints folds), and a local crenulation fabric. The metamorphic grade is mostly greenschist, based on the assemblage quartz-sericite-chlorite, but ranges to amphibolite grade in the north and south. Local andalusite and cordierite isograds have been identified and contact metamorphic aureoles are present around granodiorite-tonalite-granite intrusions.

The High Lake property features lenses (AB and D zones), pods (C, E and H zones) and other deposits farther afield (Lake zone, North zone and West zone). The largest deposit is the AB zone, which measures 200 x 600 m (Petch, 2004). It is bound to the east by High Lake and to the west is intruded by granodiorite (2613±3 Ma; Henderson et al, 1995). Other granitic intrusives in the vicinity of High Lake have been dated at 2605±5/3 Ma (Henderson et al, 1995).

Rocks in the immediate vicinity of the AB zone include felsic and less commonly intermediate flows, diabase sills and dykes, hyaloclastite, volcaniclastic rocks and carbonate-rich exhalite. Minor units of metasediment include slate, psammite, pellet and argillite. A unit of anthophyllite-magnetite-cordierite is closely associated with the ore bodies. Other alteration assemblages in the vicinity of the AB zone include silification, chloritization and sericitization. A separate assemblage of cordierite and/or quartz-sericite-chlorite porphyroblasts forms a characteristic “dalmatianite” fabric in the immediate vicinity of the ore deposits and is a useful prospecting tool (Petch, 2004).

Mineralization includes massive to semi-massive and stringer-zone pyrite, pyrrhotite, chalcopyrite, sphalerite and minor galena. Additional phases include magnetite and minor hematite. Local textures include colloform banded pyrite and vugs lined with drusy quartz, pyrite and chalcopyrite. Zoning with respect to copper and zinc is, in the case of the AB deposit, poorly developed (Petch, 2004). A Cyprus-type, bimodal, rift-like environment has been suggested as the setting for the High Lake deposits (Bleeker and Hall, 2007). Franklin et al, however, proposed a bimodal-felsic volcanogenic massive sulphide model.

**Heninga Lake (Medium)**

The dominant rock types in this central Rankin-Ennadai greenstone belt are dacites with lesser rhyolite, andesite and basalt. Pyroclastic deposits, notably dacitic lapilli tuffs with clasts up to 30 cm in diameter but mostly less than 5 cm, are prominent amongst the felsic volcanics. Minor sedimentary rocks are most commonly grey cherts which occur interbedded in the volcanic rocks. All the rocks have been metamorphosed to lower or middle greenschist facies and have suffered regional deformation, including a northeast-striking foliation associated with isoclinal folds. Later deformation includes a north-trending set of parasitic folds with flattening of volcanic clasts and axial-planar quartz veins. The general pattern is of a northeast-trending volcanic pile with younger strata located to the southeast as indicated by the facing direction of volcanic pillows (King, 1975).

Mineralization is found in exposures near the shores of Heninga Lake. The forms of mineralization include: 1) massive and medium-grained pyrite-chalcopyrite-sphalerite with lesser pyrrhotite and magnetite; 2) coarse-grained and massive pyrite-sphalerite; 3) disseminated blebs and stringers of pyrite-chalcopyrite-sphalerite, and 4) streaks and stringers of pyrite-chalcopyrite. In general, the disseminated mineralization is situated structurally above the massive type. Drilling and geophysics have delineated a zone of 850 m length within which there are separate Cu-Zn-Ag- rich and Zn-Ag-rich sections (King, 1975).
NEOARCHEAN OROGENIC GOLD

Orogenic gold deposits of Neoarchean age, all located in the Slave craton of the Northwest Territories, are considered in this section (Figure 4). The Giant mine (large) lies on the west side of Yellowknife Bay, on the north shore of Great Slave Lake. It is located in the northerly-striking Yellowknife Greenstone Belt. Giant, together with the nearby Con (also large), are the largest gold deposits of the Slave craton (followed in size by Goose Lake, Courageous Lake, George Lake, Lupin, and Colomac, all of which have been mined). Although separated by faulting, the Giant and Con mines are generally considered to be parts of a single ore deposit. The gold zones of the Yellowknife Gold project, 90 km N of the town of Yellowknife are also located in this greenstone belt. Farther afield, in the Slave craton, the Ulu deposits are located in the High Lake greenstone belt, and, to the northeast, in the High Lake belt, the gold deposits of Boston, Doris and Madrid.

Giant (Large)

Gold in this area was discovered by GSC staff who panned the Yellowknife River in 1898. Significant discoveries of lode gold were, however, not made until 1935. This led to the staking of claims on what would become the Giant Yellowknife Gold Mine and to a staking rush in 1936. Significant new discoveries were made in 1944, which precipitated another staking rush. The mine was active from 1948 to 2004 (Figure 7; Moir et al., 2006; Canam, 2006).

The basement of the Yellowstone Greenstone Belt consists of gneisses and plutonic rocks dated at 3.3 - 2.93 Ga, and supracrustal rocks of the Bell Lake Group (>2.8 Ga). Gold-bearing shear zones occur in the Yellowknife Bay Formation (2.71-2.70 Ga) and in felsic tuffs and porphyries of the Townsite Formation. Lode gold also occurs in the younger Jackson Bay Formation (breccia,

Figure 7. Yellowknife Royal Oak Giant Mine head frame. GSC 2015-110.
conglomerate, sandstone, argillite) and in the overlying Banting Group (2.67-2.66 Ga; felsic tuff and mafic volcanics). Younger intrusive rocks include: Defeat Suite granodiorite (2.63-2.62 Ga), Duckfish Granite (2.61 Ga) and Anton Complex (2.64-2.61 Ga) (Cousens et al., 2006).

In general, the Giant deposit is bound on three sides by faults and to the east by the Banting Group and Jackson Bay Formation. The Giant Mine is classified as a quartz-carbonate, shear-zone-hosted vein deposit. Schist envelops the ore zones and is closely associated with alteration as follows: Sericite-carbonate schist grades outwards to chlorite-sericite schist and finally chlorite schist. Ore-hosting schist is associated with shear zones and is also folded into antiforms and synforms. Schistose zones are known to transect bedding. The regional-scale folds are therefore not the product of simple folding of primary stratigraphic units. Four deformation phases are documented at Giant: layer-parallel fabrics (S1), east-west compression to form the main phase of deformation (S2), folding of S2 (D3), and brittle faulting of Proterozoic age (D4). The ore of the Giant Mine (and of Con Mine) is considered to be synchronous with D2 deformation and post-date the intrusive Defeat Suite (2.62 Ga). A Re-Os isochron on pyrite yields 2.59 Ga for sulphides in the Con Mine (Ootes et al., 2011).

Mineralization occurs as: 1) bands of quartz and sulphides alternating with sericite-carbonate schist; 2) sericite-carbonate schist with matrix of quartz and sulphides, and; 3) folded and boudin-aged quartz-carbonate veins. Ore minerals include: pyrite, arsenopyrite (closely associated with gold), sphalerite, chalcopyrite, stibnite, sulphosalts and pyrrhotite (Canam, 2006).

Yellowknife Gold project (Medium)
The Yellowknife greenstone belt, traceable for 50 km north from Yellowknife Bay, is bound to the east by Burwash Formation sediments. In general, the stratigraphic units strike N50°E. Graded beds indicate that tops are to the southeast. The Yellowknife Formation is part of the Kam Group, a 10 km section of tholeiitic basalt including massive, pillowed and variolitic flows, and flow breccias. There are four contained formations of which the Townsite Formation is most distinctive (dacitic flows, breccias, felsic tuff). Two intervals of pillow basalt form distinctive markers in the Yellowknife Bay Formation. The top unit in the Kam Group is the Kamex Formation: pillowed and volcaniclastic rocks with interbedded tuffs and tuffaceous sediments. N-/NW- striking gabbro dykes of multiple generations occur in the Yellowknife Greenstone Belt (Hauser et al., 2006).

The first claims were staked in 1944 on a visible gold discovery in quartz veins near Giauque Lake. Further gold discoveries were made in 1954 on a visible gold discovery in quartz veins near Giauque Lake.
1945. A shaft was sunk in 1946 followed by mine development in 1948 - 1949. Mining commenced in 1950 and by 1960 had reached a depth of 1237 m. The mine was closed in 1969. The interest of Tyhee Development Corp. in the Yellowknife Gold project dates to 2001.

The Ormsby Nicholas area is the northernmost of three mineral zones within the Yellowknife Gold project. Garnet amphibolite metavolcanic rocks within a “sea” of Burwash Formation metasediments are associated with this zone. Above the metavolcanics there is a transitional facies in which metasediments are intercalated with thin amphibolite beds. A high-strain zone referred to as the Discovery Shear Zone is associated with the mineralization. The host rocks at Ormsby are actinolite-albite-carbonate schists, commonly brecciated and derived from calc-alkalic to tholeiitic basalt. Gold occurs in folded quartz veins along with 1-10 % pyrrhotite, lesser pyrite, arsenopyrite and rare chalcopyrite, sphalerite and galena (Pratico, 2009). At Nicholas the host quartz veins are located within a shear adjacent to a small pluton of granodiorite or within Burwash Formation metasediments. The sulphide phases are the same as at Ormsby. The granodiorite also contains scheelite.

Ulu (Small)

The Ulu deposit was discovered by BHP Minerals in 1989. WPC Resources is the most recent owner. The Flood zone, the largest of 16 gold showings, strikes at 118° and dips 70-80° south. Within this zone there are 15 distinct veins, of which the largest is 500 m long and 1.5 to 18m wide and is traceable to a depth of at least 600 m. From its margins inwards it shows the following alteration assemblages: 1) biotite-titanite-tourmaline; 2) silicification (actinolite-carbonate-sericite-clinopyroxene), and 3) silicification (quartz) with arsenopyrite+K feldspar and gold. Ore zone sulphides, in descending order of significance include arsenopyrite (closely associated with gold), loellingite, pyrrhotite, pyrite, chalcopyrite, and rare sphalerite and galena (Graham and Wahl, 2011).

Hope Bay: Boston, Madrid, Doris (Medium)

Three significant gold deposits and 75 smaller showings are located in the northeastern part of the Slave craton of western Nunavut within the northerly-trending Hope Bay greenstone belt. A port is planned to provide access by way of the Northwest Passage. The potential for gold was recognized soon after the mapping of greenstones in the area in 1962. The Boston deposit was discovered in 1992 followed by Madrid in 1994 and Doris in 1995. These properties are currently worked by TMAC Resources Inc.

Volcanic rocks in the five known extrusive suites range from iron-rich tholeiitic rhyolite in the oldest succession (Flake Lake suite of rift affinity: 2716 - 2697 Ma) to calc-alkaline dacite to rhyolite in the four overlying packages (arc affinity: 2690 - 2662 Ma). These are succeeded by fluvial conglomerates containing detrital zircons, dated at 2715 - 2664 Ma and sourced from the underlying volcanic suites. These, in turn, are invaded by gabbros (2663 - 2262 Ma) and by four phases of granitoids (including tonalite, granodiorite and monzogranite) ranging in age from 2672 - 2608 Ma (Sherlock, 2012). Showings and deposits are closely associated with iron-rich tholeiites in the lower part of the volcanic succession. There is also a close association with iron carbonate alteration and pyrite-bearing quartz veins.

The Doris deposit, located in the northern part of the greenstone belt, is hosted in three quartz vein systems. The mineralogy of the veins includes quartz, sericite, chlorite, tourmaline and gold. The host rocks are overprinted by a hydrothermal assemblage including ankerite, ferroan dolomite, sericite, pyrite and other minor phases. Structural interpretation indicates that the lode gold-quartz veins were emplaced in a saddle-reef structure during D2 strain (Sherlock, 2012). The Madrid deposit, in the same part of the greenstone belt, has a different character. The gold ore occurs in sulphidized iron-rich tholeiite. Hydrothermal alteration at Madrid is associated with a quartz-carbonate stockwork and features sericite, magnesite and ankerite. The gold ores are more explicitly associated with secondary albite, paragonite, ankerite and pyrite, and with hematite alteration (Sherlock, 2012).
The Boston deposit is located in the southern part of the greenstone belt, in association with a belt of ankerite, quartz and sericite alteration that has been traced for 9 km. The ore occurs in a high strain zone in mafic pillow basalt (tholeiite and iron-rich tholeiite) that is gradational to overlying argillite and siliciclastic sediments (Sherlock, 2012).

Colomac (Medium)

The Colomac property is located 220 km NW of Yellowknife. Gold was discovered in this area by prospectors in 1938 and near Baton Lake (N of Indin Lake) in 1945. Five companies were attracted to the area and staked claims on the Colomac site. Prospecting soon led to the discovery of gold at both the East Zone and the West Zone. In 1986 Neptune Resources Corp. optioned the Colomac property and demonstrated that the deposit was amenable to open pit mining. A production decision was made in 1987 and the deposit was brought to production from 1990 to 1997. The current operator is Nighthawk Gold Corporation.

Colomac lies within the larger Indin Lake property, which has 20 known gold deposits and showings. The Colomac property is underlain by mafic to felsic volcanics and intermediate intrusives. These are bound to east and west by metasediments including argillite, greywacke and siltstone. Lenses of pyrite-pyrrhotite iron formation and sulphide-bearing argillite are also present. The volcanic rocks are intruded by a multiphase subvolcanic sill complex consisting of diorite, quartz diorite and gabbro. Part of this is identified as the Colomac Sill (2671 ±10 Ma), a differentiated quartz-albite porphyry of tonalite-trondjhemite composition with up to 2% pyrite and pyrrhotite. It is 40 - 200 m thick and traceable for 6 km (Trinder, 2013). The Colomac ore zones feature quartz veins with gold commonly associated with pyrite, chlorite, pyrrhotite, tourmaline, arsenopyrite, magnetite and sericitic alteration (Trinder, 2013).

Courageous Lake (Large)

The Courageous Lake property is located 240 km NE of Yellowknife near a winter road linking Yellowknife to Diavik and Lupin. Gold was first discovered by prospecting in the 1940s and specific deposits were found in 1944 (Tundra) and 1947 (Salmita). New deposits (the FAT and Carbonate zones) were uncovered by Noranda in 1980 and a shaft was sunk in 1988. However, results were not encouraging and the excavated gold mineralization was stockpiled at the surface. The property was purchased in 2002 by Seabridge Gold Inc (Huang et al., 2014).
The Yellowknife Supergroup within the Courageous-Mackay greenstone belt, 3 - 7 km wide and extending along strike for 70 km, is host to the mineralization. These rocks have attained mid-greenschist facies. The supergroup rocks include early tholeiites and later calc-alkaline volcanics forming a single dominant mafic to felsic cycle. Greywacke and siltstone are increasingly prominent in the upper part of the sequence. Four phases of deformation have been identified - two ductile phases producing foliation fabrics and cleavage, and two late stages of brittle faulting.

The FAT deposit consists of 13 discrete zones collectively 1900 m in length, 800 m wide and known to a depth of at least 800 m. The host rocks are felsic volcanic rocks that range from dacite to rhyolite with textures including lithic lapilli tuff, welded tuff, ash tuff and agglomerate. The deposits show sericitic, silicic, carbonate and potassic alteration. Associated sulphides include pyrite, pyrrhotite, arsenopyrite (closely associated with gold), sphalerite and chalcopyrite. An epithermal origin, involving a hydrothermal system for the gold mineralization at Courageous Lake is favoured by Seabridge Gold (Huang et al., 2014).

Lupin (Medium)

Gold was discovered in the Lupin area by Canadian Nickel Company (Canico) staff in 1960, and was mined by Echo Bay Mines Limited from 1984 to 2004. The latest owner is WPC Resources Inc.

The host unit is the Neoarchean Contwoyto Formation, a turbidite- and iron formation-dominated part of the Yellowknife Supergroup of the Slave craton. The Contwoyto Formation is succeeded by the Itchen Formation, another turbidite unit with carbonate concretions but no iron formation. There is some indication that the Contwoyto Formation may grade laterally into the volcanic flows and tuffs of the Point Lake Formation. Above this there are six plutonic suites of Archean age, all cut by Proterozoic gabbro dykes. Metamorphic isograds in the vicinity of the Lupin property are related to local granodiorite plutons and include sillimanite-in and cordierite-in isograds. Lupin lies almost entirely below the cordierite isograd. The emplacement of the ore is linked to intrusion of the granodioritic Contwoyto Batholith (2.585-2.580 Ga) and related metamorphism and hydrothermal activity in the late Neoarchean (Bullis, 1991).
The ore is hosted by Contwoyto Formation turbidites. In the immediate vicinity of the ore deposits there is a unit of greywacke and quartzite overlain by Lupin banded iron formation (Lupin BIF) which is succeeded by phyllite. The five known ore zones (West, Central, Eastern, M1 and M2) are confined to amphibolitic iron formation and the deposit is identified as being of iron-formation-hosted lode gold type (Figure 9). Types of iron formation found include silicate, sulphide and oxide (Greusebroek and Duke, 2005).

There are three deformation phases. S2/S3 include an axial planar cleavage. A major ductile shear zone is mapped at the mine (D3) and is labelled as the Lupin Deformation Zone (LDZ). Gold ores occur where the LDZ penetrative shear fabric is located on the west and central limbs of an F2 dome. Lupin BIF has been traced for 3 km and to a depth of 1500 m (Greusebroek and Duke, 2005). The host rock contains quartz-grunerite ±magnetite, where unmineralized, and the ore zones consist of hornblende-quartz-chlorite-native gold ± pyrrhotite, arsenopyrite, loellingite. In general, there are 5-30% sulphides in the mineralized zones. Quartz veins ranging from a few cm to 1m wide and traceable for up to 12m are also present. The gold can be seen, in polished sections, to be closely associated with arsenopyrite and loellingite. Accessory minerals include scheelite, chalcopyrite, chlorite, pyroxene, graphite, epidote and ilmenite. Arsenopyrite also occurs as coarsely crystalline haloes around quartz veins (Harron, 2012).

Characteristic dimensions for the ore zones are:
West zone: 220 m long x 2.5 m thick, West zone - South: 300 m x 2.0 m, Central zone: 225 m x 5 m thick. The M1 and M2 zones are largely mined out but down dip potential remains (Harron, 2012).

**George Lake and Goose Lake (Medium, Large)**

The Back River Joint Venture was established in 1982. Surface discoveries were presumably made before 1985 as drilling began in that year. These properties are currently held by Sabina Gold and Silver Corp.

Volcanic-turbidite rocks of the Yellowknife Supergroup are dominant in the Hackett River and Back River areas. These are subdivided into Hackett River (felsic to mafic volcanic), Back (felsic to intermediate volcanic), and Beechey Lake (turbidite) groups. Mineralized iron formation occurs in the Beechey Lake Group (turbidites). Facies of iron formation include oxide (magnetite-chert-grunerite) and silicate (chert-grunerite-chlorite). Both contain iron carbonate. There are indications that three folds in the iron formation represent a repetition of a single iron formation bed.

Three deposits are recognized at Goose Lake: Goose Main, Llama, and Umwelt. The following description is specific to Umwelt but to varying degrees also applies to the others. The bulk of the mineralization occurs in oxide-facies iron formation (chert-grunerite-magnetite) with lesser amounts in silicate-facies iron formation and in clastic sediments (greywacke, siltstone and mudstone). There is also a close association with quartz-carbonate veining, sulphidization, brittle faulting and folding. Mineral species include pyrite, arsenopyrite, pyrrhotite, rare chalcopyrite and free gold (Kent et al., 2013).

At George Lake, mineralization variously occurs in oxide-facies iron formation with less in silicate iron formation. There is a spatial relationship to shear zones in some deposits at George Lake, but a genetic relationship has not been proved. The gold is associated with sulphides in iron formation and with quartz veins. Minerals present include pyrrhotite, pyrite, arsenopyrite, loellingite, gold and minor chalcopyrite. The dominant amphibole is hornblende. Gold-mineralized quartz veins are also present (Kent et al., 2013).

These ore bodies are banded iron-formation hosted gold deposits with similarities to Lupin.
Nechalacho (Thor Lake) (Large)

Uranium claims were first registered on the Thor Lake property in 1970: it was then acquired by Bluemount Minerals Ltd. Niobium (Nb) and tantalum (Ta) were subsequently found in 1976 by Highwood Resources Ltd. Property work and drilling from 1976 to 1979 resulted in the discovery of Nb, Ta, Y (yttrium) and REE (rare earth elements). Active companies on the property since then have included Placer Development Ltd., Hecla Mining Co., Conwest Exploration Co. Ltd., Royal Oak Mines Ltd., Dynatec Corp., Beta Minerals Inc., and (to date) Avalon Rare Metals Inc.

The large Nechalacho deposit is located within the Blatchford Lake Complex, which is intrusive into the Slave craton north of the East Arm of Great Slave Lake (Figure 4). Named intrusive phases include (early) western lobe: peridotite, pyroxenite, layered gabbro, leucoferrodiorite, anorthosite, quartz syenite, and granite, and eastern lobe: peralkaline granite, and (late) syenite. Nepheline syenite is also present, intrusive into the Thor Lake syenite which is host to the Thor Lake deposit. Based on cross-cutting relationships, three distinct late phases of intrusion are documented in the immediate vicinity of the deposits: Grace Lake granite, Thor Lake syenite, and the Nechalacho layered suite. The age range of the Blatchford Lake Complex is 2185±5 Ma to 2175±5 Ma, whereas the ages of the Thor Lake and Nechalacho phases are 2176.8±1.6 Ma and 2164±11 Ma respectively (Mumford, 2014; Ciuculescu et al., 2013).

The Grace Lake granite is an arfvedsonite-aegirine-perthite granite with ca. 25 % quartz. It is gradational to the Thor Lake syenite which suggests that these two phases are synchronous and not intrusive into each other. The Thor Lake syenite consists variously of leucosyenite, fayalite-aegirine syenite and arfvedsonite-aegirine syenite (Mumford, 2014). The Nechalacho layered suite features nepheline and sodalite as primary rock-forming minerals. There is also a cumulate zone containing apatitic phases such as eudialyte and a so-called “basal zone” which hosts the mineral deposits (Figure 10).

Mineralization in the basal zone includes 4.6 - 9.1 % ore minerals including allanite, monazite, bastnaesite and synchysite (sources of LREE), fergusonite (for Y, HREE, Nb, Ta), ferrocolumbite (Nb) and zircon (HREE, Nb, Ta, Zr). Gangue minerals include quartz, plagioclase, microcline, biotite and iron oxides (Ciuculescu et al., 2013). The Nechalacho deposits have been compared to other peralkaline intrusions, notably: Strange Lake, Labrador; Illimaussaq in Greenland, and; Lovozero on the Kola Peninsula in Russia. These are intrusions formed by igneous differentiation, crystal fractionation and crustal assimilation. REE are concentrated in the residual liquid and the eutectic point is depressed by elevated fugacity of gases such as fluorine and carbon dioxide.
Two large Ni-Cu-PGE deposits of uncertain Paleoproterozoic age, spatially associated with the 2800 km long Snowbird Tectonic Zone (STZ), (Figure 5) are included in this group. This major tectonic break follows the boundary between the Rae and Hearne cratons in the southeast Northwest Territories (Nickel King deposit) and southwest Nunavut (Ferguson Lake deposit). The age and significance of the STZ remains controversial with interpretations including: intracontinental shear zone at 2.6 Ga; intracontinental rift at 2.55 Ga, and; a plate boundary suture at 1.9 Ga (PEG Mining Consultants Inc., 2010).

Nickel King (Large)

Canico (now Vale Inco) discovered nickel at Thye Lake in 1952. The property has been held by Strongbow Resources since 2004.

The Nickel King property is part of the Rae craton and is located 20 km NW of the Snowbird Tectonic Zone. Dominant features on the property include Neoarchean to Paleoproterozoic pelitic to psammitic paragneiss with mafic to ultramafic intrusive rocks. The psammites consist of biotite, garnet, sillimanite, quartz and feldspar. Rare meta-conglomerate preserves primary sedimentary clasts. The range of mafic to ultramafic intrusive compositions includes norite, gabbro and pyroxenite. Norite is the principal ore host. The mineralogy of the norite includes enstatite, clinopyroxene, hornblende, plagioclase and phlogopite.

The metasediments are dominated by a bedding-parallel gneissosity defined by alignment of biotite. Other features in these rocks include tight to isoclinal folds, and younger upright folds. Gabbro/norite bodies are grossly devoid of deformation fabric. However, there is petrographic evidence for strain, and there is evidence to indicate that some gabbro sills have been isoclinally folded, like the enclosing gneiss.

Five ore zones, all hosted by norite, have been identified at Nickel King. The largest, the Main Zone (2600 m in strike length), is interpreted to form a tight recumbent syncline (with upper and lower mineralized sills) which has been refolded in an open fold. The richer mineralization is net-textured to semi-massive and 2-14 m thick. Disseminated zones are mostly 60-80 m thick but range up to 120 m. The typical sulphide assemblage is pyrrhotite with inclusions of pentlandite and chalcopyrite. Grain size ranges from 0.5 - 7 mm (PEG Mining Consultants Inc., 2010). Nickel King has some resemblances to the gabbro-norite hosted Ni-Cu-PGE deposits of the Limpopo metamorphic belt of eastern Botswana, and the Las Aguilas Ni-Cu deposits in Argentina (PEG Mining Consultants Inc., 2010).

Ferguson Lake (Large)

Canadian Nickel Company Ltd. (Canico; now wholly owned by Vale Inco) discovered nickel at Ferguson Lake in 1950. The East and West zones were tested by drilling from 1950 to 1955, resulting in the discovery of significant resources to depths of 240 m. Ore zones were uncovered east and west of Ferguson Lake as well as under the lake.

The Ferguson Lake property is located in the Hearne craton southeast of the Snowbird Tectonic Zone; part of the Yathkyed greenstone belt (~2.71 to 2.66 Ga) consisting of Archean supracrustal and intrusive rocks metamorphosed to upper amphibolite facies, Paleoproterozoic plutons and younger dykes. The supracrustal rocks are identified as mafic volcanics, chert iron formation, lesser intermediate to felsic volcanics and clastic metasediments. The Archean intrusives include granodiorite, quartz monzonite, diorite and gabbro (2.66-2.63 Ga). Host rocks
in the vicinity of the Ferguson Lake property include east- to northeast-striking sills (?) of amphibolites and gabbro, sulphide- oxide- and silicate-banded iron formation, quartz-feldspar-biotite gneiss and paragneiss, all intruded by Archean tonalite, granite and pegmatite. The host gabbro/hornblendite displays compositional layering including pyroxenite to leucogabbro (in the West zone) and mesocratic to leucocratic gabbro and anorthosite (in the East zone) (Campos-Alvarez et al, 2012). Three generations of deformation are recognized. Antiforms and synforms have developed in gneisses and metagabbros. Mineralization, hosted by hornblendite, is located in the south limb of a recumbent, doubly-plunging synform, subsequently modified by shear zones and faults.

The host of the magmatic Ni-Cu-PGE mineralization is gabbro and hornblendite. This same body is 10 - 600 m thick and traceable from East to Central to West zones, a distance of 12 km. Better grades occur in lenses, pods and stringers (two to tens of metres thick) of massive to semi-massive ore consisting of 80-90% pyrrhotite, lesser chalcopyrite, pyrite and pentlandite. Platinum group minerals identified include two palladium tellurides, three palladium bismuthinides, and palladium and platinum arsenides. Other textures include brecciated ores (gabbro clasts in a sulphide matrix), and net-textured ores noted in stringer and fracture-filling zones. The Ferguson Lake ores have been identified as tholeiitic intrusion-hosted nickel-copper ores (associated with differentiated intrusions) (Clow et al., 2011).

ULTRAMAFIC-HOSTED NICKEL-COPPER PGE

Large Ni-Cu-PGE deposits of commercial significance are located in Paleoproterozoic ultramafic rocks in the Cape Smith belt of northern Ungava (province of Quebec; Figure 11). The Cape Smith belt (2.04 - 1.86 Ga) has been interpreted as a stack of southerly transported klippen consisting of quartzite, semipelite, ironstone and gabbro-peridotite in the lower part (Povungnituk Group) and basalt, and gabbro-peridotite in the upper part (Chukotat Group). These rocks accumulated on a long-lived north-facing rifted margin.

The age of the mineralized ultramafic rocks is 1.883 Ga. The host rock is the volcano-sedimentary Povungnituk Group consisting of the Nituk Fm. at the base (conglomerate, sandstone, siltstone, silicate iron formation); the Beauparlant Fm. (tholeiitic basalt, phyllite); the Cecilia Fm. (alkaline basalt, andesite, mafic rhyolite, pyroclastics, sediments), and the Nuvilic Fm. (deep marine sediments including exhalative sulphides). The Nuvilic Formation is intruded by the Lac Esker Suite ultramafic intrusions, which host the West Raglan deposits. The overlying Chukotat Group is largely volcanic (komatiitic to tholeiitic basalts) and is coeval with the Lac Esker Suite (also intrusive into the Chukotat Group). The Lac Esker Suite consists mainly of subvolcanic ultramafic sills, predominantly composed of wehrlite but ranging from dunite to peridotite and pyroxenite. Individual intrusions are <150 m thick and traceable for 1 - 10 km. Overlying this there is a mostly volcanic island-arc accretionary complex, the Watts Group.

The Ni-Cu-PGE deposits are located primarily in Chukotat peridotite (Lesher, 2007). The Raglan, West Raglan and Nunavik deposits are described below. Neoarchean nickel mineralization also occurs in ultramafic rocks in the vicinity of Rankin Inlet, Kivalliq region, southwestern Nunavut.

Raglan (Large)

Low-grade showings were first discovered in 1898, in the western part of the Cape Smith Belt by A.P. Low of the Geological Survey of Canada. Sporadic exploration has taken place in the Cape Smith Belt and in the Raglan deposit area since the 1930s with the first high-grade showings being discovered by prospectors in 1956. Mining was first attempted by New Quebec
Raglan Nickel Mines from the 1960s to 1970s but was then abandoned due to low metal prices. Mining was renewed from December 1997 when Falconbridge Ltd. Xstrata acquired Falconbridge Ltd. in 2006. A merger in 2013 led to formation of the current mine owner, Glencore Xstrata.

The dominant mineralization at Raglan is footwall-contact type composed of disseminated, net-textured and massive pyrrhotite, pentlandite and chalcopyrite, contained in over 140 lenses located from the surface to a depth of 750 m (Figure 12). Lens tonnages range from 0.01 to 5.2 Mt, averaging 0.2 Mt. The basal layer in each lens is typically massive, overlain by net-textured ore that grades into disseminated mineralization. Massive to semi-massive brecciated ore also occurs and consists of a mixture of footwall sediments and ultramafic rock. Another feature is injection of sulphides into host rock sediments. The “up” direction at the time of sulphide emplacement was to the north, in sill (or dyke) rocks that are considered to have been subvertical. Ore lenses typically occur in embayments at the base of ultramafic sheets. Similarly, canoe-shaped intrusions feature “keels” of sulphide, tens to

Figure 11. Simplified geology of northern Quebec (Cape Smith belt) and adjacent southeastern Baffin Island together with deposits of nickel-copper-PGE and kimberlite diamond.
hundreds of metres in length. The ore zones occur in linear clusters that collectively define the channel in which they occur (Desharnais, et al., 2014).

**West Raglan (Large)**

The West Raglan property (owners: True North Nickel Inc. and Royal Nickel Corp) is located in the western part of the Cape Smith Belt, which consists mainly of 1.88 Ga-old magmatic rocks associated with the Chukotat Large Igneous Province. Three types of ore lens have been identified: contact, hanging-wall and vein-type mineralization. Seven zones have been delineated at West Raglan. In general, the mineralization occurs at contacts between ultramafic intrusive sills (Desharnais, et al., 2014).

**Nunavik (Large)**

The Nunavik Nickel mine is located 80 km W of Kangiqsujuaq (Wakeham Bay) and 30 km ESE of the Raglan Mine in the Nunavik region of northern Quebec. Disseminated sulphides were first discovered in the area in 1957. The current owner is Jilin Jien Nickel Industry Co. Ltd.

The Ni-Cu deposits in the Cape Smith belt lie in the Raglan Trend in the north and in the South Raglan Trend located 20 km to the south. Although in many ways similar, there is a contrast in Ni/Cu ratios: 3:1 in the Raglan trend and closer to 1:1 in South Raglan. The host rock for the deposits in the Nunavik Nickel property, located in the South Raglan Trend, is the Expo Intrusive Suite, a 500 m wide intrusive complex, consisting of gabbro dykes, ultramafic bodies, dunite pipes and sill-like peridotites. These have been emplaced into the Povungnituk Group and are collectively traceable for 50 km in an east-northeasterly direction. Host-rock ultramafic bodies have a dyke- or channel-like (or trough-like) geometry, individually 100-200 m wide and traceable along strike for several kilometres. In general, the sulphides occur in the basal part of various ultramafic sheets with massive sulphide lenses at the base grading up into net-textured (25-75%) sulphides and a wider halo of less significant disseminated sulphides (Armstrong et al., 2010).

Seven mineralized zones have been defined within the Expo Intrusive Suite. The following description applies to one of these, the Allammaq zone. The deposit is hosted in peridotite that varies in attitude from a dip of 20°N to near vertical and has been emplaced into metabasalt and metasediment of the Beauparlant Formation (part of the Povungnituk Group). Drilling indicates the presence of four distinct units within the host intrusion: 1) pyroxenite-peridotite, 2) hornblende gabbro; 3) sparsely mineralized pyroxenite-peridotite, and 4) heavily mineralized pyroxenite. Sulphides occur as disseminations, blebs, net-textured ores, and vein-like massive sulphides.
In descending order of significance the sulphide mineralization includes pyrrhotite, pentlandite (major carrier of cobalt), chalcopyrite and accessory galena, sphalerite, cubanite and cobaltite. Platinum group minerals include sudburyite (lead antimonide), PGE tellurides including michenerite (a palladium bismuth telluride), merenskyite (a palladium telluride), moncheite (a platinum telluride), kotulskite (a palladium telluride), and sperrylite (platinum arsenide); gold is carried in electrum (Armstrong et al., 2010).

Rankin Inlet (Small)
The deposit is located on the west shore of Hudson Bay near the community of the same name (Figure 5). It was discovered in 1928, grade and tonnage figures were obtained in 1929 and the deposit was mined by North Rankin Nickel Mines Ltd. from 1957 to 1962. The deposit is located in the Rankin Inlet Greenstone Belt which is subdivided into a lower volcanic cycle, a clastic sedimentary horizon and an upper volcanic cycle. The lower volcanic cycle consists of a mixed assemblage of volcano-sedimentary rocks dated at 2663±3 Ma (Tella et al., 1996). The upper cycle consists of thick, pillowed flows and minor interflow sediments with a date of 2629±14 Ma (Tella et al., 1996).

The mineralized sill has been emplaced along the contact between greywacke sediments, tuffs and quartzites (below) and upper cycle volcanics (above). It consists of pyroxenite and serpentine-rich peridotite. Mineralization occurs in depressions along the base of the host-rock sill. Massive sulphides at the base grade upwards into net-textured and disseminated ores. Metallic phases include pyrrhotite, pentlandite, chalcopyrite, magnetite, pyrite, gersdorffite, violarite and marcasite.

PALEOPROTEROZOIC OROGENIC GOLD

Gold deposits of presumed or established Paleoproterozoic age are hosted in Neoarchean iron formation in the Rae and Hearne cratons of southwestern Nunavut (Figure 5). The Cullaton Lake (not described), Meadowbank, Meliadine (large) and Three Bluffs deposits are considered here. The Meadowbank deposits are located 70 km N of Baker Lake community in the Rae craton. The Three Bluffs deposit is also located N of Baker Lake. The Meliadine property is located in the Hearne craton 25 km NW of Rankin Inlet on the west shore of Hudson Bay: it is connected to Rankin Inlet by all-weather road.

Meadowbank (Large)
The Meadowbank deposit, currently owned by Agnico-Eagle Mines Ltd, is primarily hosted by iron formation of the Neoarchean Woodburn Lake Group. The stratigraphy of the Group includes quartzite (at the base), conglomerate, ultramafic metavolcanics (talc-chlorite-amphibole) and intermediate volcaniclastic rocks with wacke, mudstone and Algoma-type iron formation (Sherlock et al., 2004). Felsic volcanics in the Woodburn Lake Group are dated to 2.735 - 2.71 Ga ((Sherlock et al., 2004; Ruel et al., 2012). Detrital zircons in quartzite are dated at 3.0 - 2.81 Ga. There is widespread intrusion of granitoids dated to 2.62 - 2.60 Ga. The metamorphic grade ranges up to granulite facies regionally, but the grade near the deposit is upper greenschist to lower amphibolite facies. Geologic relationships and geochronology indicate that the gold mineralization was introduced during D2 deformation (isoclinal folds, axial planar foliation and shear zones) in the Paleoproterozoic (1.9 - 1.8 Ga; Sherlock et al., 2004). All these units are overlain by Paleoproterozoic strata of the Baker Lake Basin.

Meadowbank has four significant ore zones: Portage, Goose, Vault and Bay. The Portage and Vault zones are currently being mined (Agnico-Eagle, 2016). The largest, Portage, is 1.85 km x 100-230 m wide and 6-8 m thick. Goose is 750
m x 500 m wide and 3-20 m thick. Vault is 1100 m long and 8-18 m thick (Ruel et al., 2012). The Portage and Goose deposits are identified as iron-formation-hosted lode gold type. Vault is a lode-gold deposit of disseminated/replacement type. The primary mineralogy of the iron-formation host rock is magnetite, quartz and amphibole, in layers 0.2 to 5.0 cm thick. The deposits are associated with widespread sulphide and native-gold replacement of magnetite. Sulphide minerals include disseminated to semi-massive pyrrhotite and pyrite with minor chalcopyrite and arsenopyrite. Gold also occurs in quartz-sulphide veins.

Meliadine (Large)

The Meliadine deposits (Agnico Eagle Mines Ltd.) are located within the Rankin Inlet greenstone belt, part of the Hearne craton. Supracrustal rocks encountered in the greenstone belt include mafic volcanics, felsic pyroclastic rocks, sediments and gabbro sills. The gold deposits and showings are closely associated with the Meliadine trend, a W-NW-trending belt of supracrustal rocks that includes a major structure, the Pyke Break. The Pyke Break is a several kilometres wide high-strain zone that is spatially associated with the known gold deposits. The Meliadine trend includes seven ore zones: Tiriganiaq, Discovery, Normeg, Wesmeg, F zone, Pump and Wolf. Ore emplacement is considered to have occurred during the third deformation phase in the Paleoproterozoic, but is postdated by crenulation.

The stratigraphic setting of the Tiriganiaq deposit features the following units: turbidites (Sam Formation), iron formation with chert, chloritic mudstone and greywacke (Upper Oxide Formation), siltstone (Tiriganiaq Formation) and basalt, gabbro dykes and interflow sediments (Wesmag Formation). The oldest gabbros feed mafic volcanics while the youngest are post-tectonic and are associated with ore emplacement. The ore horizons are spatially associated with the Upper Oxide Formation but in other deposits (Wesmag, Normeg) they extend into the Wesmag Formation. The gold mineralization is, in general, associated with shearing and quartz veining during the Paleoproterozoic (Orosirian) Trans-Hudson orogeny (1.9-1.8 Ga). Re-Os ages on arsenopyrite yield ages of 2.27 and 1.90 Ga (Lawley et al., 2015). This suggests phases of mineralization that are coincident with the Arrowsmith and early Trans-Hudson orogenies.

Gold-hosting quartz occurs as veins several metres thick but ranging downwards in size to erratic quartz stringers and stockwork. Typical vein mineralogy includes quartz, and quartz-ankerite. In contrast, quartz-calcite veins are barren. Primary sulphides introduced during mineralization occur in iron formation and argillite as wispy discontinuous laminae of pyrrhotite,
pyrite, chalcopyrite and arsenopyrite (Figures 13, 14). Late-stage sulphides include galena and sphalerite. In general, sulphides are not a reliable prerequisite for high gold grades. Ore-deposit types include orogenic-greenstone and banded iron-formation-hosted gold mineralization (Larouche et al., 2015).

Three Bluffs (Medium)
The Three Bluffs gold deposit (North Country Gold Corp) is also hosted by Archean iron formation. The age of the mineralization is Paleoproterozoic. These iron formations lie within the Committee Bay belt (2.73 - 2.69 Ga), which consists of komatiite-quartzite-iron formation that has been metamorphosed to amphibolite and granulite facies and intruded by diverse granitoids (2.61 - 2.59 Ga). Younger granitoids include granite, granodiorite and monzogranite (1.82 Ga) (Davies et al., 2010).

Host rocks for uranium and silver arsenide vein deposits lie within the Great Bear magmatic zone (1.87 to 1.84 Ga), part of the western domain of Wopmay orogen, a magmatic and collisional belt that brought together the Slave craton located to the east and the Hottah terrane to the west. Included in this category of deposits are the Eldorado, Echo Bay and Terra Silver mines near Port Radium on Great Bear Lake (Figure 15). The belt also includes iron oxide copper gold deposits (IOCG: NICO, Sue-Dianne).

Port Radium (Medium)
Silver and uranium deposits near Port Radium are located along the eastern shore of Great Bear Lake, Northwest Territories (Figure 16). These include the former Eldorado uranium and silver mine and the Echo Bay silver deposit. The first record of iron, copper, uranium and cobalt near Echo Bay was made by J.M. Bell of the Geological Survey of Canada in 1900. High-grade pitchblende and silver ores were discovered in 1930 by the prospector, Gilbert LaBine. Mineral production began at Eldorado in 1933, but closed briefly in 1940, reopening in 1942 to supply uranium to the war effort. Production was maintained, more or less continuously, to 1982. The Echo Bay deposit was brought into production for silver and copper from 1964 to 1975.

Host rocks in the area include the Port Radium Formation of the LaBine Group (sandstone with lesser carbonate), Cobalt porphyry and andesite of the Surprise Lake Member. These are intruded by quartz monzonite and granite of the Great Bear magmatic zone, and by diabase of the Cleaver and Western Channel swarms. Studies have shown that two types of granite are present. One of these is associated with high heat flow and may be a source for uranium in the Port Radium area (Somarin and Mumin, 2012).

Ore deposit metals are associated with steep northeast-striking faults in andesitic deposits of the Great Bear magmatic zone. The earliest
mineralizations, sparse chalcopyrite and pitchblende in quartz, are seen to heal these brittle faults. These are then cut by Cleaver diabase dykes dated at 1.74 Ga (Gandhi et al., 2013). Post-Cleaver fault reactivation led to a total of five stages of mineralization as follows.

Stage 1: pitchblende with chlorite, fluorite, carbonates; Stage 2: arsenides, nickel sulphides, native silver, native bismuth. These first two stages precede 1.66 Ga Narakay Islands volcanics.

Stage 3: polymetallic sulphides, silver tellurides; Stage 4: carbonates, native bismuth, minor native silver. These stages are intruded by Western Channel Diabase at 1.59 Ga. The fifth and final stage, post-Western Channel diabase, produced additional native silver and bismuth (Gandhi et al., 2013).

**Terra Silver (Small)**

The Terra Silver property is located on the southeast side of Great Bear Lake in the Northwest Territories. Although there were
staking rushes for nickel, uranium and silver in the area from 1929, little was accomplished until the 1960s. High-grade silver and copper intersections were reported by Terra Mines Ltd from the drilling campaign of 1967. Mining occurred from 1969 - 1985. Historic production at Terra Silver includes high-grade silver plus cobalt, bismuth and nickel arsenides hosted in a gangue of quartz, albite and carbonate. The veins are located in the Terra Formation of the Labine Group which is identified as caldera-fill of fluvial and lacustrine sediments. Ten percent of the host sediments contain banded and disseminated sulphides including pyrite, pyrrhotite, chalcopyrite, arsenides, native silver and bismuth. Native silver-bearing pods in the host veins are enveloped by an alteration assemblage that includes quartz, hematite, chlorite and carbonate. Minor minerals include skutterudite, safflorite, rammelsbergite, matildite and sphalerite (Webb, 2003).

Vein uranium is commercially significant at Rayrock in the southern Great Bear magmatic zone and at Lac Cinquante in the Hearne craton W of Hudson Bay in southwestern Nunavut.

**Rayrock (Small)**

The Rayrock uranium mine is located at the southern end of the Great Bear magmatic zone, 40 kilometres N of the north arm of Great Slave Lake. Pitchblende was first identified in the area by Geological Survey of Canada staff in 1934. Intensive prospecting in the 1950s led to the discovery of the Rayrock deposit which was mined from 1957 to 1959. The mineralization at Rayrock features uranium and copper in giant quartz veins in brittle fractures that post-date Great Bear magmatic activity. The ore-bearing veins are up to 60 m wide and traceable for 5 km. The host rock is gneissic granodiorite that has been mylonitized and altered. Alteration assemblages up to 10 m wide include silicification, chloritization, epidotization and hematization. Most of the quartz veins are barren but
some contain hematite, copper sulphides, pyrite and pitchblende. Study of the paragenesis indicates early pitchblende followed by bornite, chalcocite, covellite and hematite. Lead isotopic ages on pitchblende indicate an age of 517+80 Ma, which suggests a genetic association with transgressive Cambrian strata (Gandhi, 1994).

**Lac Cinquanté (Medium)**

The discovery of polymetallic showings and a uranium deposit at Lac Cinquanté dates to the 1960s. Independent of their uranium potential many of the showings are enriched in copper and silver. Most uranium-related work ended in 1981. Renewed interest in uranium and in iron oxide-copper-gold potential arose in 2007. The property is held by Kivalliq Energy Corporation. The nearest settlement to Lac Cinquanté is Baker Lake, 225 km to the northeast (Figure 5).

The Lac Cinquanté deposit is located in the Angikuni Greenstone Belt of the Chesterfield Block, part of the Hearne craton. It is close to the unconformity below the transtensional Paleoproterozoic Baker Lake Basin (1.84 - 1.79 Ga). Other Paleoproterozoic depocentres include the Wharton (1.76 - 1.74 Ga) and Thelon (1.72 - 1.54 Ga) basins. Contemporaneous intrusives include the Hudson (1.85 - 1.81 Ga) and Nueltin (1.76 - 1.75 Ga) granitoids (Bridge et al, 2013).

The deposit is hosted by Archean graphitic-tuffaceous metasediments. Typical strata of the overlying Baker Lake Basin near Lac Cinquanté include a discontinuous fault-related basal breccia overlain by conglomerate with granitoid clasts and an arkose that grades laterally to trachyte (Christopher Island Formation) (Bridge et al, 2013). The deposit is fracture-controlled, brecciated and associated with alteration phases including veins of hematite, chlorite, quartz, carbonate and pitchblende. Ore shoots lie within the plane of the host tuff and have been traced along strike for 3.5 km, and to a depth of 400 m. The mineralogy of the host includes pyrite, chalcopyrite, molybdenite, galena, sphalerite, pitchblende, coffinite and a variety of other uranium phases (Figure 17; Dufresne et al., 2013). The pitchblende has been dated to 1.83 Ga (U-Pb), coinciding with the Hudson granitoid suite and deposition of the Baker Lake Group (Bridge et al, 2013).

The deposit type has been compared to the basement- and structure-hosted uranium vein type, not unlike that found in the similar age Beaverlodge district of Saskatchewan (1.9 to

Figure 17. Lac Cinquanté Main Zone: foliated graphite chlorite sulphide tuff; pervasive hematite alteration and quartz pitchblende veinlets and breccia. GSC 2015-113.
1.8 Ga). Two theories have been proposed for the origin of Lac Cinquante. The first is that the uranium is derived by leaching from Baker Lake Basin and the Hudson granite and concentrated in basement graphitic rocks which acted as a reductant; the second is that the uranium derives entirely from Baker Lake strata and was concentrated in a hydrothermal system associated with emplacement of the Hudson granite (Bridge et al, 2013).

UNCONFORMITY URANIUM

Three deposits in this category are closely associated with the unconformity below Paleoproterozoic strata of the Thelon Basin in Northwest Territories and Nunavut (Figure 5). These include the Boomerang Lake deposit (Small, not described) in southeast Northwest Territories, the Kiggavik deposit W of Baker Lake in southwestern Nunavut, and the Andrew Lake deposit (not described) near Kiggavik, also in Nunavut.

Kiggavik-Andrew Lake trend (Medium)

The Kiggavik deposits are located 80 km W of Baker Lake in the Kivalliq region of Nunavut. Exploration in the area has been ongoing since the 1970s. The Kiggavik Main Zone (KMZ) and two other deposits were discovered in 1974 by lake/water geochemical methods and by airborne radiometrics. The project consists of five uranium deposits, three at Kiggavik and two at a separate location (Sissons site). Four of the deposits are to be mined by open pit, and one by underground methods. Other discoveries were subsequently made at Andrew Lake and Endgrid 15-17 km SW of Kiggavik. In 1993, controlling interest in the property passed from Urangesellschaft to COGEMA Inc. (Areva) who undertook a pre-feasibility study in 1997.

Figure 18. Kiggavik Main Zone: showing vermiform texture (irregular roll front like feature) and graded density of uraninite in remobilized uranium zones. GSC 2015-112.
The deposits lie just outside the Paleoproterozoic Aberdeen Sub-basin of the northeastern Thelon Basin. The deposits lie in basement and are separated from the Thelon Basin by the Thelon Fault. The basement consists of highly deformed but weakly metamorphosed Neoarchean to early Paleoproterozoic metasediments, metavolcanics and intrusive rocks. These were intruded by Hudson Suite granite and by ultrapotassic mafic to felsic dykes, all dated at 1.83 Ga. The basal unit of the overlying Dubwnt Supergroup is the Baker Lake Group: conglomerate, sandstone, ultrapotassic volcanic rocks, syenite, microsyenite and lamprophyre.

The middle unit of the Dubwnt Supergroup is the Wharton Group; associated with the 1.75 Ga Kivalliq Igneous Suite (KIS: volcanic and epiclastic rocks, Nueltin granite, anorthositic gabbro, and diabase in two swarms (Robinson et al., 2014). This is overlain by the Thelon Formation pink sandstone.

The host rock at Kiggavik is Neoarchean quartzofeldspathic metagreywacke with minor iron formation and metapelitic, unconformably overlain by 2.6 Ga rhyolite (ore host) and early Paleoproterozoic quartzite (barren) (Robinson et al., 2014). The three ore zones discovered to date are: the East (EZ), Main (KMZ) and Centre (CZ) zones. The largest is KMZ, hosted by greywacke and KIS granite. Ore is also enclosed by alteration and features desilification and conversion of feldspar to sericite and illite. The ore minerals include uraninite, coffinite and minor uranophane as fine disseminations, veinlets parallel to foliation and fracture fillings (Figure 18). Other minerals in trace quantity include galena, copper sulphides, gold and electrum, molybdenite, bismuth minerals, iron oxides and others (Robinson et al., 2014).

SANDSTONE-HOSTED URANIUM

Amer Lake (Medium)

The Amer Lake property is located 150 km N of Baker Lake and 70 km N of the Meadowbank gold deposit in southwestern Nunavut (Figure 5). Interest in the area dates to 1969, when an airborne radiometric survey by Aquitaine led to the discovery of uranium mineralization in sandstone. The Main and Faucon zones were outlined by drilling in 1970. Significant lateral continuity of mineralization was proven by Uranelz in 1978. The property has been held by Adamera Minerals since October 2015.

The geological context includes Archean basement (>2.7 Ga) of the Rae craton infolded with supracrustal rocks of the Woodburn Lake Group (2.72 Ga), all intruded by granitic and dioritic rocks (2.6 Ga). These are overlain unconformably by the Paleoproterozoic Amer Group (2.45-2.1 Ga) which underwent deformation, intrusion and regional metamorphism associated with the trans-Hudson orogeny (1.91-1.80 Ga). Overlying this orogen there are generally unmetamorphosed Paleoproterozoic sediments of the Baker Lake (1.85-1.79 Ga), Wharton Group (1.75 Ga) and Thelon (1.72 Ga) basins (Armitage, 2009).

The uranium mineralization at Amer Lake is associated with exposures of the Amer Group E of Thelon Basin. Host formations consist of arkose, feldspathic sandstone, quartz arenite, mudstone and minor dolostone. The target for exploration is sandstone-hosted uranium as for other deposits of this type located in Australia, southwestern US, South Africa and Kazakhstan. Ore minerals of the Main zone at Amer Lake include uraninite, lesser brannerite and a secondary phase, uranophane (Armitage, 2009).
These deposits include the NICO and Sue Dianne deposits of the Great Bear batholith, part of the Wopmay orogen W of the Slave craton (Figure 15). Also considered in this group are the Wernecke breccia deposits (Pagisteel) located in the Ogilvie Mountains of northern Yukon, and the younger Minto (large) and Carmacks copper deposits located in the Canadian Cordillera of central Yukon (Figure 44).

**NICO (Medium)**

The NICO property is located in the vicinity of Mazenod Lake, 160 km NW of Yellowknife. Mineralization was discovered in the area in the 1930s. The first property work on known Co-Bi-Cu arsenide showings was by New Athona Mines Ltd, from 1968 to 1970. Drilling uncovered the additional occurrence of gold. New discoveries were made by Fortune Minerals Ltd who acquired NICO in 1994, and recognized favourable similarities to the Olympic Dam deposit in Australia. Delineation drilling and pre-feasibility studies were carried out in 1998 and 1999, and bulk sampling in 2007.

The NICO deposit occurs in the Great Bear magmatic zone (GBMZ), part of the Wopmay orogen. The GBMZ extends northwards from Great Slave Lake to eastern Great Bear Lake and consists of calc-alkaline volcanic and plutonic rocks (1.88 to 1.84 Ga; Goad et al., 2000). It occupies the orogenic suture between the Coronation platform margin of the Slave craton (to the east) and the Hottah terrane (to the west). The GBMZ consists of low-titanium oxide and high-alumina calc-alkaline volcanic and plutonic rocks. Volcanic rocks of the Faber Group, varying from felsic to intermediate, are prominent in the southern GBMZ. Rhodacite, ignimbrite, flows, tuffs, breccias and volcaniclastic rocks are found and are associated with granodiorite, monzogranite, rapakivi granite and feldspar porphyry (Burgess et al., 2014).

The strata of the Coronation platform margin include shelf and slope deposits of the Snare, Akaicheto and Epworth groups. The Snare Group (now referred to locally as Treasure Lake Group and as cover of the Hottah terrane) is exposed on the NICO property: arenite, dolomite, siltstone, shale. The host rock of the NICO deposit is the Treasure Lake Group (TLG) that is brecciated and altered by iron- and potassium metasomatism, and is close to the sub-Faber Group unconformity. The contacts between metasediments and granites are mylonitic. The proximal host is biotite-amphibole-magnetite schist in the Treasure Lake Group. Other TLG rocks include subarkosic wacke, arenite, minor siltstone and carbonate. The Faber Group on the property (1851+18/-16 Ma) includes rhyolite, rhodacite, sub-volcanic intrusive equivalents and volcaniclastics. GBMZ granites are found nearby and the TLG is hornfelsed along the contacts. Breccias along the contact with Faber Group volcanics are interpreted as hydrothermal diatreme breccia bodies. The breccias have TLG and felsite clasts in a matrix of iron oxides, biotite, amphibole, chlorite and potassium feldspar (Burgess et al., 2014).

The NICO deposit is mostly contained within magnetite ironstone, schist and subarkosic
wacke containing 3-10% sulphides (Figure 19). The ore minerals include cobaltite, cobaltian arsenopyrite, bismuthinite, chalcopyrite, a gold bismuth telluride and native gold. Gangue minerals include pyrite, pyrrhotite, magnetite, hematite and silicates (Burgess et al., 2014). NICO is classified as an iron-oxide-copper-gold deposit of either Olympic Dam or Cloncurry sub-type.

**Sue-Dianne (Small)**

Early mapping in the region was carried out in the period 1936-1939 by Geological Survey of Canada staff. A radiometric survey by the GSC uncovered significant anomalies north of Mazenod Lake and resulted in staking of the Sue and Dianne claims in 1974. New work included extensive staking, geology, geophysics and airborne gravity. Drilling by Fortune Minerals commenced in 1997.

The Sue-Dianne breccia complex features an outer zone of intense shearing and alteration, with cross-cutting veins, stockwork, breccia, silicification and epidote “flooding”. The veins are parallel to the ENE-striking Dianne Lake fault. Zoning around the deposit includes: 1) quartz-epidote veining, stockwork and gouge; 2) potassic altered and iron-rich breccia, sparsely mineralized with malachite, pitchblende and uranium oxides, 3) mineralized diatreme breccia with a matrix of magnetite and hematite (Figure 20), and 4) a breccia cap rock. The deposit occurs in a diatreme breccia body 600 m long, 500 m wide and at least 350 m deep; the deposit itself is 450 m x 300 m x 350 m deep. Copper sulphides include chalcopyrite, bornite, chalcocite and covellite. Silver and gold are associated with bornite and chalcopyrite but are not seen as independent phases (Hennessey and Puritch, 2008).

**Pagisteel (Small)**

Iron-oxide-copper±gold±uranium±cobalt (IOCG) mineralization occurs in breccia bodies within the Wernecke Supergroup of the Ogilvie Mountains of northern Yukon. The Wernecke Supergroup consists of three units: the Fairchild Lake Group, Quartet and Gillespie Lake groups.

Wernecke breccia bodies occur throughout the Wernecke Supergroup but preferentially in the Fairchild Lake Group. Individual breccia bodies range in dimensions up to several kilometres. The breccia matrix consists of rock fragments and hydrothermal precipitates: feldspar, carbonate (calcite, dolomite, ankerite), quartz and more locally hematite, magnetite, chalcopyrite, biotite, muscovite, barite, fluorite, minor tourmaline and actinolite (Figure 21; Hunt et al., 2010). In total there are 65 known breccia occurrences, all associated with copper, iron and uranium mineralization. Associated minerals of economic interest include chalcopyrite, magnetite, hematite, pitchblende and brannerite (an Fe-Ti oxide containing uranium and cerium) (Hunt et al., 2010). The origin of the Wernecke breccia remains uncertain but it may be related to overpressured conditions, and transport and dissolution of evaporites from the upper Fairchild Lake Group.
Minto (Large)

The deposit (owner: Capstone Mining Corp) is accessible on the Klondike Highway via Carmacks and Minto to Minto Landing on the Yukon River. The Minto deposit lies within the Yukon-Tanana terrane, host of the Carmacks Copper Belt and of several intrusion-related Cu-Au hydrothermal systems. Major features include metaigneous and metasedimentary rocks of Permian age lying on pre-Late Devonian basement. The Minto deposit is located in the Granite Mountain batholith which intrudes the Yukon-Tanana terrane. The batholith is overlain by Late Cretaceous Tantalus Formation sediments, and by andesite and basalt of the Carmacks Group, also Late Cretaceous. The batholith is flanked to the E of Minto by mafic volcanic rocks thought to be of Triassic age.

The mineralization occurs in rocks that have a strongly developed fabric, all of which are considered to be variant facies of the primary granodiorite. The possibility that the deformed rocks are metasedimentary or metavolcanic rafts has been ruled out. The primary mineralization consists of chalcopyrite, bornite, chalcocite, minor pyrite, magnetite and accessory covellite, hessite, native gold and electrum. Textures include disseminations and sulphide veinlets parallel to the foliation. Grades increase in zones of intense folding. There are also massive and semi-massive sulphide zones which obliterate primary host-rock fabrics. In the Minto Main deposit there is a zonation from bornite-rich (up to 8%) in the west to chalcopyrite-rich but lower grade mineralisation in the east. Precious-metal grades are higher in the bornite zone. These zonation trends are also observed in other parts of the deposit (Mercer et al., 2012). The supergene assemblage, present at depths of 30 - 90 m, includes chalcocite, malachite, azurite and rare native copper. Supergene gangue phases include limonite, hematite and clay-altered feldspar.

Carmacks (Medium)

The Carmacks deposit is located in the Dawson Range, 220 km N of Whitehorse. Geologic components of the region include the Whitehorse Trough, the Yukon Crystalline Terrane and the Yukon Cataclasite Terrane. The rocks of the Whitehorse Trough, E of the Carmacks property, include the Late Triassic Povoas Formation (intermediate to mafic volcanics), carbonate reefs of the Jurassic Lewes River Group, and Early Jurassic greywacke, shale and conglomerate of the Laberge Group. The Yukon Cataclasite Terrane consists of the Granite Mountain granodiorite with screens and pendants of strongly foliated gneiss. These latter rocks are the host of the Carmacks copper deposit. The Yukon Crystalline Terrane, exposed SW of the deposit, consists of early Paleozoic mica schist, quartzite, marble and amphibolite intruded by Jurassic and Cretaceous granite and syenite. Younger strata, not included in the trough and terrane successions, include the Late Cretaceous Carmacks Group and the Mount Nansen volcanics (Kent et al., 2014).

The deposit is underlain by intrusive and meta-intrusive rocks of the Granite Mountain intrusion, ranging in composition from granodiorite to diorite and with textures varying from porphyritic to massive and foliated. Rock types associated with the ore zones include biotite gneiss, quartzofeldspathic gneiss, amphibolite and biotite schist. Siliceous ore is also found. Aplite and pegmatite dykes up to 3 m wide postdate the mineralization.

The No.1 zone is 700 m in strike length, and open to depth below 450 m. The supergene zone extends to 250 m and has been the principal focus of exploration. Drill holes were historically terminated at the base of the metal oxide zone. The mineralogy of the oxidized ore includes malachite, cuprite, azurite, tenorite and minor amounts of covellite, digenite and djurlite (Figure 22). Primary metallic phases include magnetite, rare molybdenite, native gold, native bismuth, bismuthinite, arsenopyrite, pyrite and pyrrhotite in a gangue of carbonate (Kent et al., 2014).
Stratabound deposits of the Canadian Cordillera of Yukon and Northwest Territories lie in outer shelf and deep-water basin strata of Neoproterozoic to Carboniferous age (Figure 23). These include iron deposits of Neoproterozoic and Jurassic ages, Mississippi Valley-type (MVT) deposits hosted by Mesoproterozoic to Devonian shelf carbonates, sedimentary exhalative deposits (SEDEX) found in deep-water sediments of Cambrian to Devonian age, and volcanogenic massive sulphide deposits that formed in Paleoproterozoic to Mississippian volcanic rocks.
The (very large) Neoproterozoic Crest iron deposit is located in the headwaters area of the Snake and Bonnet Plume rivers of the Mackenzie Mountains of northern Yukon. This is a remote area, 580 km N of Whitehorse and close to the border with the Northwest Territories. This group also includes the (medium-sized) Alto deposit, of presumed Jurassic age (Figure 23).

**Crest (Snake River) (Very large)**

The deposits were discovered in 1961 by Standard Oil Company geologists who identified 10 - 30 m thick sections of jasper hematite iron formation. Crest Exploration acquired the deposit in 1962. Subsequent reconnaissance determined that the deposit might contain 15 billion tons of iron ore. However, evaluation showed that the deposit could not compete with iron from Australian sources. Evaluations by Kaiser Engineers in 1976, 1991 and 1998 again showed that the deposits were sub-economic because of their remote location. The present owner is Crest Exploration Ltd, a subsidiary of Chevron Canada Resources.

The deposit is located in a westerly-trending portion of the Mackenzie Mountains within a package of folded, mostly carbonate-dominated, Upper Cambrian to Upper Devonian strata. The deposit is bound to the south by Precambrian and lower Paleozoic strata. The Snake River iron formation occurs in the lower part of the Rapitan Group of Neoproterozoic age. The host clastic rocks are conglomerate and siltstone ≥2000 m thick, maroon-coloured in the lower 610-915 m and highly ferruginous, with iron formations present at several levels in the lower 305 m of the formation.

The Crest deposit consists of fine-grained specular hematite with alternating jasper-rich bands. It has been traced for 51.5 km. Iron formation that is economically significant occurs up to 305 m above the sub-Rapitan contact. This zone attains a maximum thickness of 120 m of which 85-105 m is iron formation. Interlayered lithologies include hematite, dolostone, ankeritic carbonate, shale, sandstone, shaly conglomerate, and conglomerate. Shaly conglomerate is the most significant lithology between the iron formation layers. Types of iron formation include nodular, banded, and irregular intergrowths of hematite and jasper (Figures 24, 25). The average iron content is 43%, the main impurity being apatite (McBean, 2006).

**Alto (Medium)**

The iron deposits at Alto were discovered in 1973 and staked in 1975-76 by a joint venture including Inexco Mining Co., Amoco Canada Petroleum Co. Ltd., Arrow Inter-America and Husky Oil Ltd. Rio Alto performed road building, trenching and mapping in 1983. The ground was restaked by Eagle Plains Resources Ltd. in 1996. Oolitic magnetite iron formation is exposed in a 46 m thick bed at the contact between the Permian Jungle Creek Formation and the Jurassic-Lower Cretaceous Kingak Formation. The exposure has a strike length of 366 m (Anonymous, 2013a).
Figure 24. Crest deposit: Boulder of banded jasper and specular hematite. Nodular hematitic layer at the top. (Yukon Geological Survey).

Figure 25. Crest deposit: Outcrop of banded jasper and specular hematite (Yukon Geological Survey).
Carbonate hosted zinc-lead deposits range from Mesoproterozoic to Devonian in age and are located in the Canadian Arctic Islands (Polaris and Nanisivik deposits, both large), along the northern edge of the Western Canada Sedimentary Basin (Pine Point deposit, large) and in the Mackenzie Mountains (Blende, the large Gayna River deposit, Goz Creek (not described) and the large Prairie Creek deposit).

**Blende (Medium)**

The property is located 115 km N of Mayo, Yukon. Although there was active staking and prospecting in the area from the 1920s, mineralization was not discovered on the Blende property until 1961 which led to further staking by Cyprus Anvil Mining Corp. in 1975. Blind Creek Resources acquired an option on the Blende property in 2006 and a 100% interest in 2008 from the previous owners, Eagle Plains Resources Ltd.

The stratigraphy of the property is dominated by the Gillespie Lake Group, part of the Mesoproterozoic Wernecke Supergroup. The lower part of the Gillespie Lake Group consists of greenish-grey to brownish-orange laminated dolomitic siltstone with a well-developed cleavage. The central units of the Gillespie Lake Group consist mostly of orange-tan dolomitic siltstone. These strata host the mineralization on the property. Columnar stromatolites, 3-15 cm wide, rare oolitic layers, and intraclast conglomerates are also found. The highest part of the Gillespie Lake Group consists of thick bedded dolostone weathering to a reddish-orange colour.

The ore zone host rock is the Mesoproterozoic Gillespie Group, in the upper part of the Wernecke Supergroup and in close proximity to a presumed ESE-striking fault of Mesoproterozoic age. Mineral occurrences follow the fault zone for 6000 m. They occur as sulphides in discordant veins, in fault-related breccias and as high grade concentrations instromatolitic units. The vein mineralogy is mostly sphalerite and galena with lesser pyrite.Anglesite, covellite and smithsonite are found in weathered materials. Zoning in the deposit ranges from spotty copper- and silver-bearing (chalcopyrite, freibergite, tetrahedrite) in the lower part in the west through lead-rich (galena and anglesite) in the middle and upper parts. Zinc enrichment (as sphalerite altering to smithsonite) occurs in the eastern part of the deposit (Price, 2011). Pb/Pb isotope model ages of mineralization are in the range of 1490 - 1430 Ma which indicates an early Mesoproterozoic age (Moroskat et al., 2015). The mineralization has been identified as fault-hosted breccia of carbonate-hosted Irish type.

**Gayna River (Large)**

This property is located in the headwaters of Gayna River, Northwest Territories, 186 km W of Norman Wells and 298 km E of Mayo, Yukon. The property was first staked by Rio Tinto Canadian Exploration Ltd (Rio Canex) in 1974. Property evaluation and drill hole testing occurred from 1975 to 1979. The claims were allowed to lapse in the 1980s. The property was restaked in 2000 and is now 100% owned by Eagle Plains Resources.

Zinc-lead mineralization of Mississippi Valley-type (MVT) is located in the Little Dal Group, part of the Neoproterozoic Mackenzie Mountains Supergroup. Although mineralization occurs in many units, the better showings are in the Grainstone formation. The mineralogy of the showings is sphalerite (with high gallium and germanium contents) and galena with a silver association. The claim block lies at the hinge of an anticline. The bedding dips 5-100 to SW or NE. Dolomitization and mineralization appear
to be unrelated to local thrust faults or other structures (Hewton, 1982).

The Grainstone formation has four members of which the first and third from the base are host for sphalerite and galena mineralization (Figure 26). Both of these units contain cavity-filling calc spar, dolospar and barite. The lower host features these phases in bedding-parallel veins while the upper host unit has breccia matrix spar. Sphalerite is more concentrated in the lower host unit whereas the upper unit has more showings.

Alteration in the lower and upper host units of the Grainstone formation takes the form of early-generation fine-grained dolomite that is grey-black, grey-white or red-brown. A white variety of dolospar with barite and fluorite fills early dolomitization porosity and is closely associated with sphalerite and galena. Most mineralization-related dolomite at Gayna River has obliterated primary carbonate textures, is present as pore- and cavity-filling cement, and acts as a substrate for sulphide deposition.

Nanisivik (Large)

Nanisivik was mined from 1976 to 2002. The nearest settlement is Arctic Bay which is accessed from the mine and airport by an all weather gravel road. Concentrate was removed by ship from a purpose-built port in the immediate vicinity of the mine.

The Nanisivik deposit is located in Mesoproterozoic rocks of the Borden Basin on northern Baffin Island. Borden Basin is interpreted to have formed by rifting from approximately 1270 Ma, with episodic extensional faulting occurring during basin evolution. Sedimentation was terminated by inversion at ca. 1000 Ma. The host of the Nanisivik deposit is the Society Cliffs Formation: a dolostone (>1000 m thick) featuring deep-water laminates and deep water carbonate mounds in western Borden Basin. In the eastern Borden Basin the Society Cliffs is dolostone forming peritidal cycles, in total 250 m thick. Pb-Pb dating indicates an age of 1199 Ma (Dewing et al., 2007).

Local features of the ore body include a laminated dolostone host rock. The orebody occurs in a horst block; part of a zone of east-striking normal faults. The ore body is 3 km long, 100-200 m wide and 10 - 30 m thick. In addition, there is an ore zone keel 65 m deep and 5-30 m wide. The ore is mostly pyrite with zones rich in sphalerite, galena and dolospar. The age of the orebody is unresolved but determinations range from 1095 Ma (paleomagnetic dating of hydrothermally altered dolomite) to 461 Ma (dating of feldspathic alteration). The latter age is unlikely as the deposit is cut by a Franklin dyke (dated at 723 Ma) (Dewing et al., 2007).

Early genetic models emphasized a void-filling karst-related origin for the Nanisivik deposit. Later models emphasize staged replacement of host rocks involving a gas cap trapped beneath the overlying Victor Bay Formation shale, with only minor, initial karst-related porosity in the Society Cliffs.

Pine Point (Large)

The Pine Point properties are located 800 km N of Edmonton near the south shore of Great Slave Lake. Occurrences of lead and zinc at Pine Point were first discovered here in 1899 by Robert Bell of the Geological Survey of Canada. Shipment of high grade ore by Cominco Ltd was initiated in 1965. Production continued until mine closure in the late 1980s. The property is now held by Tamerlane Ventures Inc.

The Pine Point district lies within the Western Canada Sedimentary Basin, which in this area comprises 350 - 600 m of sediments ranging
from Ordovician to Devonian in age and limited to the east by the Canadian Shield. In the Middle Devonian, a carbonate barrier (the Presqu’ile Reef) formed along a SW-trending basement ridge with open marine conditions to the north and restricted back reef facies (the Elk Point lagoon) to the south. These restricted conditions led to deposition of gypsum, anhydrite and rock salt. Carbonate deposition continued into the Late Devonian with a phase of uplift and karsting marking the end of barrier reef development. In modern nomenclature, the Presqu’ile facies is thought to be a diagenetic alteration superimposed on a variety of primary reef-associated depositional units.

The mineralization is genetically linked to mineralizing brines and diagenetic alteration forming in karst-induced porosity in the host rock. The host stratigraphy is limited to six named Middle Devonian units to a maximum of 200 m of section. In total there are 97 known deposits within three northwest ore trends, distributed over a strike length of 68 km and a width of 6 km. Forty eight of these deposits were mined by Cominco Ltd. before 1990. The deposits have the form of vertical pipes (karst chimneys), and tabular bodies that lie along former subsurface stream channels (Siega and Gann, 2014).

The mineralization occurs as sphalerite, galena, marcasite and pyrite and as a replacement of karst-fill sediments, breccia, open-cavity fillings (Figure 27) and as peripheral disseminations. Better grades are associated with karst structure and replacive Presqu’ile dolomite. Sphalerite occurs in colloform masses. Galena is present in a nested form inside sphalerite. Other ore-related phases include marcasite, pyrite, minor pyrrhotite, celestite, barite, gypsum, anhydrite and fluorite. Bitumen and pyro-bitumen are also found, particularly in trap settings above the ore bodies. Hydrogen sulphide gas has also been reported (Siega and Gann, 2014). The Pine Point deposits are classic Mississippi Valley-type lead-zinc deposits, possibly the best example of its kind in Canada.

**Polaris (Large)**

The Polaris zinc-lead district spans an area of 450 km (N-S) by 130 km (E-W) coinciding with the Late Silurian-Early Devonian Cornwallis Fold Belt and the Boothia Uplift in the central Canadian Arctic Islands. Within this area there are eighty zinc, lead and some copper showings and the Polaris deposit on western Little Cornwallis Island.

Figure 27. Pine Point: typical botryoidal textured sulphides; layered sphalerite (dark brown and white) is overlain by coarse grained galena (steel blue). Sulphides were precipitated in a cavity, beginning at lower right. The last deposited sulphide was galena. Coin is 1.5 cm in diameter. GSC 1995-214.
The surface showing at Polaris was discovered in 1960. A gravity anomaly was drilled in 1970, which led to the discovery of the deposit, which was mined from 1982 until reserves were exhausted in 2002. The Polaris deposit occurs in the upper part of a Cambrian to Upper Ordovician carbonate shelf succession that notably includes evaporites in the Lower and Middle Ordovician. The host rock is the upper Thumb Mountain Formation of Upper Ordovician age. Other small showings are found in carbonate units as high in the stratigraphy as the Middle Devonian.

Specific host rocks within the upper Thumb Mountain Formation include, from the base: 1) burrow-mottled skeletal wackestone with chert nodules; 2) wackestone with algal colonies; 3) burrow-mottled skeletal wackestone; 4) macrofossil-rich fossiliferous wackestone, and; 5) argillaceous nodular wackestone. The Thumb Mountain Formation is succeeded by green shale with limestone interbeds (Irene Bay Formation), and by black shale with lime mudstone, chert and siltstone (Cape Phillips Formation).

The dimensions of the Polaris deposit are 800 m x 300 m x 20-100 m thick. There are two distinct parts to the deposit: the Panhandle zone (two ore types, in the upper part of the deposit) and the Keel zone (three ore types, in the lower part). The Panhandle (P1) ore type includes massive, carbonate replacement, breccia-fill and vein sulphides. The P2 ore consists of a vein network with veins up to 1m thick of sphalerite, marcasite and galena. Below this, in the Keel zone is K3 ore, consisting of a vein complex with disseminated sulphides. Deeper still is K2 ore, which is mostly fracture- and vein fill with minor replacement ore and breccia. Deepest in the deposit is K1 ore, comprising low-grade fracture and vein fill (Dewing et al., 2007a).

In general, the mineralization features colloform sphalerite, also occurring in disseminated form and as aggregates (Figures 28, 29). Galena occurs in dendritic and skeletal forms and in polycrystalline veins. Other minor phases include pyrobitumen, barite, spar calcite and pyrite. The age of the Polaris ore body is 366±15 Ma (Rb/Str). This coincides with the Ellesmerian Orogeny, which brought sedimentation in the Franklinian basin to a close (Dewing et al., 2007a).

Figure 28. Polaris mine: sphalerite-galena vein showing banded ore on either side and crystalline galena and sphalerite in the centre. GSC 2015-117.

Figure 29. Polaris mine: marcasite-rich ore showing collapsed fragments of banded sphalerite. GSC 2015-118.

Prairie Creek (Large)

The Prairie Creek deposit is located 500 km W of Yellowknife in the Mackenzie Mountains. The property was discovered by a trapper in 1928 and was first staked in 1956. The period of exploration up to 1970 included extensive surface drilling and underground testing.

A decision to put the deposit into production was made in 1980. A winter road was completed from the Liard River Highway, and a mill and concentrator were moved to the site between 1980 and 1982. At this point Cadillac, the principal operator, went bankrupt and the assets were taken over by Procan Exploration Co.Ltd. By 2004 Canadian Zinc Corporation had earned a 100% ownership in the property.
The stratigraphy on the property ranges from Ordovician to Devonian in age and includes the Whittaker, Road River (graptolitic dolomite), Cadillac (grey siltstone and shale) and Arnica (black dolomite and limestone) formations. The Whittaker Formation, the host of the ore deposit, has been divided into nine members. The compositions of these include massive dolostone, bioclastic dolostone (two units), mottled dolostone (ore host rock), chert nodular dolostone (three units) and interbedded chert and dolostone.

The mineralization on the property is diverse and includes: a 16 km long mineralized quartz vein system, in part linked to mineralized strata-bound zones, other vein systems extending 4 km to the N and 10 km to the S, subsurface mineralization referred to as the Main Zone, subsurface stockwork and stratabound mineralization, and Mississippi Valley type (MVT) mineralization of which there are six named showings in the northern part of the property.

The vein mineralization contains galena and sphalerite with lesser pyrite and tetrahedrite-tennantite in a matrix of quartz, calcite and dolomite. The largest vein has a probable length of 2.1 km. In other localities vein mineralization occupies tension fractures in competent strata such as the lower Road River and Whittaker formations. Some of these have an “en echelon” geometry. Stratabound mineralization, discovered by drilling, has been traced over a distance of 3 km. It consists of massive sphalerite, coarse galena, disseminated to massive pyrite, but little or no copper or sulphosalts. It occurs mainly in mottled dolomite of the Whittaker Formation. MVT mineralization occurs as colloform sphalerite with pyrite, marcasite and minor galena. It is associated with open cavities in biothermal reefs of the Root River Formation (above the Road River Formation only in the northern part of the property)( Paradis, 2007; Shannon et al., 2012).

To summarize, the main ore deposit types include hydrothermal veins, Mississippi Valley-type deposits (like Pine Point) and deposits of the carbonate-hosted Irish type.

SEDIMENTARY EXHALATIVE (SEDEX) DEPOSITS

SEDEX deposits are an important resource in the Yukon (Figure 23). These include the large deposits in the Anvil area (Early Cambrian Faro, DY, Grum, Swim, Vangorda), Clear Lake (Devonian Mississippian), the potentially very large Howard’s Pass (Early Silurian), Mel (Cambrian or Ordovician), and the large Tom and Jason deposits (Late Devonian).

Anvil (Large)

The Anvil mining camp is located 200 km NE of Whitehorse near the town of Faro (Figure 30). The Faro deposit is one of five significant deposits located in the Anvil Mining Camp of the central Yukon. The others are the Grum, Vangorda, Dy and Swim deposits. The Vangorda deposit was the first to be discovered by conventional prospecting, in 1953. Systematic exploration in the area (by Prospectors Airways Co. Ltd) began in 1956. The property changed hands several times until the Faro claims were established in 1964 by Dynasty Exploration Ltd which entered into a joint venture with Cyprus Exploration Ltd. in 1965. The Faro orebody was discovered in June 1965. Mining began in 1969 and continued until the ore was exhausted in 1991. Production was entirely shifted to the Vangorda deposit in 1992. Production ceased in 1993 due to the bankruptcy of the parent company (Curragh Resources Ltd).

The host rocks include the Lower Cambrian Mount Mye Formation overlain by the Cambrian to Ordovician Vangorda Formation. The Mount
Mye Formation consists of non-calcarous phyl­lite and schist. The Vangorda Formation consists of calcareous phyllite and impure carbonate. The five known deposits are associated with a 150 m thick interval that straddles these two forma­tions. The sulphide lenses have been traced later­ally into a carbonaceous pelite unit which has been identified as a submarine exhale layer produced through hydrothermal venting at the sea floor. The Faro deposit is close to the mid-Cretaceous Anvil batholiths, leading to meta­morphism of the host Mount Mye Formation to biotite- andalusite-muscovite schist. The occurrence of pyrrhotite in the Faro deposit has also been attributed to metamorphism.

Sulphide minerals, in decreasing order of abun­dance, include pyrite, sphalerite, galena, pyrr­hotite, chalcopyrite and marcasite. Less sig­nificant minerals include barite with traces of tetrahedrite, bournonite and arsenopyrite. Zoning is present in the Faro deposit and has been recognized as cyclical. The idealized cycle begins with ribbon-banded carbonaceous pyritic quartzites that give way to pyritic quartz­ites, siliceous pyritic sulphides, massive pyritic sulphides and baritic massive pyritic sulphides. In general, the upper part is baritic and higher grade; the lower and outer parts are lower grade, carbonaceous and quartzose. Alteration identified in the hanging wall and footwall includes “bleaching”, silification and quartz­muscovite-plagioclase alteration (Pigage, 1991). Deformation of the Faro deposit is associated with the upper limb of a Z-shaped fold. There are also extension faults that have down dropped the central part of the deposit.

The Anvil deposits are identified as sedimentary exhalative type. Characteristic features include layering of sulphides, fine-scale interbedding of sulphides with unmineralized sediments, limited stratigraphic range, and pre-metamorphic and pre-tectonic ore deposition (Pigage, 1991).

Clear Lake (Medium)

The property was first staked in 1965 by Conwest Exploration Company Ltd. at the time of discov­ery of the Faro deposit. Exploration expanded considerably with the Macmillan Joint Venture from 1975 to 1980, and then with Getty Canadian Metals Ltd (including drilling) to 1983. Renewed drilling was carried out from 1991 to 1996.

The Clear Lake deposit is considered to be a sedi­mentary exhalative deposit consisting mostly of pyrite but including a sizeable component of lead and zinc. The deposit occurs in carbonaceous argillite, siltstone, chert and tuff of the Devonian.
to Mississippian Earn Group. The property straddles the Tintina Fault with Lower Cambrian phyllite of the Mount Mye Formation to the N along with Cambrian to Ordovician phyllite and limestone of the Vangorda Formation. South of the fault the host rock Earn Group overlies Ordovician to Lower Devonian shale of the Road River Formation. The deposit has a significant gravity and EM anomaly and adjacent to this is a small lake with anomalous bottom sediment zinc values. The deposit is 1000 m long and up to 120 m wide. The sulphides are laminated and include frambooidal pyrite, interlayered with tuffaceous rocks in the stratigraphic footwall. This interval also includes concretions of galena, sphalerite, barite, siderite and calcite. Silicification is prominent in both footwall and hanging wall locations, and massive barite forms a hanging wall cap. Metal ratios for the deposit indicate a syngenetic origin related to Devonian rifting. This is also indicated by worm-tube fossils partly replaced by sphalerite (Anonymous, 2014b).

Howard’s Pass
(Large, potentially Very Large)

The Howard’s Pass property is located along the Yukon-Northwest Territories border 350 km NE of Whitehorse: most of the claims are located in the Yukon.

The Howard’s Pass deposits were discovered in 1972 during follow-up of a 1971 stream sediment reconnaissance program. This led to a significant staking rush. Anomalies of zinc, lead and cadmium were subsequently found in soil surveys. Geophysical methods proved to be ineffective. 218 drill holes were collectively drilled from 1973 to 1981 and also in 2000. Bulk testing on the XY deposit occurred in 1980 and 1981. The property is currently held by Chihong Canada Mining Ltd.

The Howard’s Pass deposit is located in the eastern part of the Selwyn Basin, a deep-water basin of Ordovician to Devonian age in which shale, chert and deep-water carbonate are predominant rocks. In general, the lower part of the basin, of Early Ordovician to Early Devonian age, has mudstone, siltstone and carbonates whereas the upper part has mudstone and siltstone with sandstone and some conglomerate. Compressive deformation of Mesozoic age has produced folds and thrust sheets across the basin.

The oldest unit in the vicinity of Howard’s Pass is the Neoproterozoic to Early Cambrian Grit Unit. This is overlain by a massive limestone (lower member) of the Rabbitkettle Formation and is succeeded by a (upper member) wavy banded limestone. Above this is the Howard’s Pass Formation, the lowest of four formations of the Road River Group. The Howard’s Pass Formation hosts all the ore deposits of the Howard’s Pass region. Typical rocks include, from the base: pyritic siliceous mudstone; calcareous mudstone; cherty mudstone; the mid-Llandovery Active Member (containing all the known zinc and lead mineralization), and uppermost, siliceous mudstone. Above the Road River Group is the Earn Group (Late Silurian to Devonian) consisting of siliceous mudstone, mudstone, siltstone and greywacke (O’Donnell, 2009).

The following account focuses on the Active Member, host of the known ore deposits. The Active Member ranges from 0-60 m in thickness and contains nine intercalated facies: grey limestone; graded limestone; calcareous mudstone; thin bedded mudstone; cherty mudstone; rhythmite; thin bedded cherty mudstone; whitish grey lead-zinc mudstone, and; grey chert. The lead-zinc mudstone is a laminated cherty unit containing up to 70% sulphides and consisting of quartz, sphalerite, galena, minor pyrite and only local calcite (Figures 31, 32). Trace constituents include chalcopyrite, tetrahedrite, molybdenite, pyrrhotite, polydymite, millerite and gersdorffite.
Depositional features include lamination, carbonate nodules, sulphide replacement of microfossils, frambooidal pyrite and bedding truncation surfaces (Jonassen & Goodfellow, 1986). The Active Member has been identified over a strike length of 37.5 km within which there are 15 deposits. The mineralized horizon is generally 20 - 30 m thick and is mineralogically consistent over the entire property. However, higher grades and coarser grain sizes have been encountered in the XY, Don and Anniv zones (O’Donnell, 2009). Syndepositional structure includes slump folds, dilational veins, quartz pressure shadows, sulphide “diapirs”, fluid escape structures and local pull-apart basins within which sulphidic sediments are concentrated (Jonassen and Goodfellow, 1986).

The Howard’s Pass deposit is classified as vent-distal sedimentary exhalative (SEDEX) type.

**Mel (Medium)**

The Mel property is located 80 km ENE of Watson Lake in the southeast Yukon. The property is currently held by SilverRange Resources Ltd. The Mel deposit occurs in Cambrian and Ordovician strata consisting of a footwall crypto-grained limestone and a hanging-wall argillite, which grades upwards to wavy banded limestone. To the north on the property there are volcanic flows and volcaniclastic sediments above the crypto-grained limestone. Mineral occurrences are present at four locations. However, only the Mel Main zone has been drilled to any extent and it is on this that known resources are based. This zone is 800 m in length and up to 21.7m thick. It consists of coarse-grained red sphalerite and galena formed in a coarse matrix of barite. Pyrite forms < 2% of total sulphides. Sphalerite also occurs as fracture fillings and irregular masses in sheared and sericitized mudstone. Minor amounts of chalcopyrite and tetrahedrite are associated with a stockwork in the cryptograined limestone (Miller and Wright, 1984; King, 1995).

Geological relationships suggest that the Mel Main zone, the Jeri zone and Mel East zone are all part of a common mineralized horizon repeated and exposed to the surface by thrusting and N-S-trending folding. The deposit type is identified as sedimentary exhalative. The coarse-grained nature of the ore, however, suggests crystallization of metallogenic brines under a thin mudstone blanket. Lead isotope ratios, though, indicate a Devonian age and therefore an epigenetic origin (Godwin et al., 1988).

**Tom and Jason (Large)**

The property is accessible via the Canol Road from Ross River. The Tom prospect was first staked by Hudson Bay Exploration Development Company Ltd (HudBay) in 1951. Surface evaluation and some drilling were carried out in 1951-1953, and again in 1967 - 1968. Underground evaluation began in 1970 including bulk testing and metallurgy. Claims were staked on the Jason property in 1974 and in 1980 this was taken over by Aber Resources Ltd. who began joint studies with HudBay at this time. A feasibility study was completed in 1986 which indicated a mine life of 15 years for the two combined deposits. There was additional drilling resulting from a Cominco option (from 1988) before this lapsed in 1992. Currently both properties are 100% owned by HudBay.

The Tom and Jason deposits are located along the eastern margin of the largely deep-water facies Selwyn Basin. Major rock packages in the vicinity of the deposits include (from the base) Neoproterozoic elastic rocks and shales; Lower Cambrian to Middle Devonian shelf carbonates; Ordovician to Devonian shales, cherts and minor limestone of the Road River formation and;
Devonian to Mississippian cherty shale and turbidites of the Earn Group (Rennie, 2007). The deposits lie in a structural domain featuring turbidites of the Lower Earn Group. Host rocks include chert pebble conglomerate, breccia, sandstone, siltstone and shale. Barite horizons up to 30 m thick are found in these rocks. Structural features include tight folding and steep reverse faulting. Some faulting has been attributed to syndepositional extensional deformation. Strata on the Tom property include pyritic mudstone with sand layers, siltstone, barite, and Pb-Zn-Ag-bearing sulphides. Local structure is interpreted as a graben bound by horst blocks. Local facies adjacent to these structures are conglomerate and diamictite. Hydrothermal vent facies include ankerite and quartz veins containing pyrite, chalcopyrite and galena.

Sulphide-barite mineralization occurs as laminae within sediment host rocks (Figure 33) and collectively is found in stratiform lenses up to 40 m thick and traceable for up to 1200 m along strike. In total there are three mineralized zones at Tom. Economic minerals include fine to coarse sphalerite, galena and minor chalcopyrite with accessory barite, pyrite, pyrrhotite, quartz, and iron carbonates. The mineralization is separately identified as Grey facies (pink sphalerite, galena, pyrite and grey barite) or as Black facies (black mudstone, cream sphalerite, galena, and pyrite). Collectively these two facies comprise the bulk of the mineralization (Rennie, 2007).

**VOLCANOGENIC MASSIVE SULPHIDE DEPOSITS (VMS)**

VMS deposits (Figure 23) of Paleozoic age have been discovered most notably in the Devonian and Mississippian strata of the Finlayson Lake area of the central Yukon (medium to large Fyre Lake, Kudz Ze Kayah, Wolverine, Ice and GP4F) but also at Hart River (not described), Hasselberg (large: Devonian-Mississippian) and Marg (Devonian-Mississippian).

**Finlayson Lake (Medium to Large)**

The Finlayson Lake district covers an area of 300 km x 50 km within the Yukon-Tanana terrane south of Ross River and northwest of Watson Lake, Yukon. Host rocks are early Mississippian to early Permian age consisting of mafic and felsic volcanic rocks, phyllite, chert, volcanic-derived sandstone and some limestone, conglomerate and diamictite in the upper part. There are also plutonic rocks that are coeval with the volcanic units. Metamorphism is mostly upper greenschist to lower amphibolite facies. Franklin et al. (2005) have identified the deposits of the Finlayson Lake district as bimodal-felsic type volcanogenic massive sulphides (VMS). However, the geological contrast between deposits would suggest a more complex genetic history. The district features five significant deposits, briefly described here, and numerous showings.

The Fyre Lake deposit (medium) is 160 km NW of Watson Lake. The host rocks include metavolcanic chlorite phyllite, quartz muscovite phyllite and sedimentary carbonaceous phyllite.
The East zone consists of massive pyrite and lesser pyrrhotite, chalcopyrite, sphalerite and magnetite. The West zone consists of: magnetite, pyrite, and chalcopyrite massive sulphide grading to pyrite and chalcopyrite, and to pyrrhotite farthest west. A chert-pyrite-chalcopyrite-magnetite layer is considered to be an exhalite. The footwall shows chlorite alteration and a stringer zone. Weak hanging wall alteration is also present (Peter et al., 2007). The deposit is classified as Besshi-type mineralization in a forearc setting or as a pelitic-mafic VMS in the classification scheme of Franklin et al. (2005).

The host rocks for the Kudz Ze Kayah deposit (large) are felsic volcanic and volcaniclastic rocks overlain by argillite and mafic volcanics. The main zone is 500 m long, 400 m deep and 18-34 m thick. The mineralization consists of pyrite, sphalerite, pyrrhotite, galena and chalcopyrite, with traces of arsenopyrite, sulphosalts and electrum. Gangue minerals include chlorite, magnetite, quartz, iron carbonate, sericite, albite and barite. Footwall alteration includes silica, albite, sericite and chlorite. There is also a footwall stringer zone (Peter et al., 2007). The deposit is classified as a Kuroko-type VMS in a back-arc basin or rift setting, or as a bimodal-felsic type VMS (Franklin et al., 2005).

Four km from Kudz Ze Kayah, the host of the GP4F deposit is Late Devonian to Early Mississippian felsic volcanic flows and volcaniclastic rocks cut by mafic dykes. The deposit is a massive sulphide lens 200 m long by 350 m down dip and ≤ 3.2 m thick. The ore consists of banded "buckshot" (pebbly) pyrite in a matrix of sphalerite, pyrrhotite and galena with local magnetite and traces of chalcopyrite. Gangue minerals include quartz, feldspar, chlorite, carbonate, sericite and biotite. In addition there is footwall sericite alteration (Peter et al., 2007). The deposit is classified as a Kuroko-type VMS, or as a bimodal-felsic type VMS (Franklin et al., 2005).

The host rocks of the Wolverine deposit (large) are Tournaissian argillite, volcaniclastic rocks, magnetite iron formation and a carbonate exhalite. The metamorphic grade is middle greenschist. The deposit dimensions are 750 m x 16m max. thickness: the deposit is open down dip. The ore types include: 1) layered massive sulphides; 2) semi-massive replacement ore, and; 3) stringer sulphide veins. The massive sulphide type includes pyrite, sphalerite, minor pyrrhotite, chalcopyrite and galena with rare antimonides, native gold and electrum (Figure 34). The ore is generally recrystallized due to deformation. However, local colloform textures and frambooidal pyrite are also present. The deposit is zoned, from copper-rich in the lower part to zinc- and lead-rich on the fringes. Footwall alteration assemblages include carbonate, silica, chlorite- and sericite alteration. There is a footwall stringer zone and weak hanging wall alteration (Peter et al., 2007). The deposit is classified as a volcanic-sediment-hosted massive sulphide (like the deposits in the Bathurst camp,
New Brunswick), or as a siliciclastic-felsic VMS (Franklin et al., 2005).

The host rocks of the Ice deposit (small) are massive, pillowed and brecciated basalts interlayered with chert, minor greywacke and carbonaceous mudstone. The deposit is 28 m thick with a copper-rich basal zone underlain by a sulphide stringer zone featuring pyrite, quartz, chalcopyrite and hematite. The sulphides are overlain by hematitic chert and, above this, by basalt. Footwall chlorite alteration is accompanied by a footwall stringer zone (Peter et al., 2007). The deposit is classified as a Cyprus-type massive sulphide. This deposit is identified as a mafic VMS (Franklin et al., 2005).

Hasselberg (Large)

The Hasselberg property is located SW of the Tintina Fault in the Pelly Mountains SE of Ross River. Newmont Exploration of Canada Ltd discovered the original showing in 1955 and staked claims in 1966. The property changed hands several times before being optioned from YGC Resources Ltd. by Atna Resources in 1995. The deposit lies within the Late Devonian - Early Mississippian Pelly Mountains volcanic belt, part of the Pelly-Cassiar Platform. The volcanic belt overlies cliff-forming carbonates, siltstones and shales of Silurian to Middle Devonian age and is overthrusted by sandstone, grit, argillite and massive carbonates of the Ordovician to Devonian Road River Formation and Earn Group-equivalent strata. The Pelly Mountains volcanic belt consists of volcaniclastic strata, lapilli tuff, argillite and lesser trachyte flows, sills and dykes. Bedded barite and massive sulphides are interbedded with these strata (Anonymous, 2014f).

The Hasselberg deposit consists of two massive sulphide lenses collectively up to 1200 m wide and extending to a depth of 500 m: they are underlain by andesitic tuffs, and overlain by rhyolite tuff and porphyry. Potassium feldspar-sericite-clay-carbonate alteration is pervasive. The upper part of the deposit features lenticular barite and disseminated pyrite, sphalerite and galena. Drilling in 1997 and 1998 indicated the possibility that the deposit, with stringer mineralization above and exhalite below, lies on an overturned fold limb. In general the deposit is 3 - 5 m thick and consists of pyrite with banded amber sphalerite and galena, or botryoidal sphalerite and galena in a matrix of Fe-Mg carbonate and lesser barite (Anonymous, 2014f).

Marg (Medium)

The Marg property is located 140 km E of Dawson near the settlement of Keno City in the central Yukon: it was first staked in response to Geological Survey of Canada stream sediment geochemical survey results. The claims were allowed to lapse when no obvious mineralization was found. Renewed interest led to drilling in 1988 which uncovered extensive volcanogenic massive sulphide mineralization. The latest owner is Golden Predator Mining Corp. in a joint venture with Minquest Ltd.

The property lies near the northern margin of the Selwyn Basin within a panel of southeast dipping strata that are intensely sheared, penetratively deformed and metamorphosed to lower- and middle greenschist facies. The property lies between two significant thrusts and features a Paleozoic metamorphic succession with up to three phases of deformation: lineation and isoclinal folding (D1), upright to isoclinal folds with steep cleavage (D2) and open to tight southwest-vergent folds (D3). The Earn Group host rocks (Devonian-Mississippian) consist of tuffs and greywacke with interbedded black shale and sulphide mineralization. The dominant rock type in the sulphide zone is quartz chlorite schist with variable carbonate content. This is thought to be a hydrothermally altered ash tuff and or pyroclastic rock.

The Marg deposit is classified as a volcanogenic massive sulphide (VMS) deposit and shares some features with other Yukon VMS prospects including Kudz Ze Kayah, and Wolverine in the Finlayson Lake area. It is hosted by felsic volcanic rocks and in this respect is similar to the Kuroko deposits of Japan and Noranda, Quebec. The mineralization at Marg occurs in multiple sheets traceable over 1400 m along strike and to a depth of 700 m. The sulphide minerals include pyrite, sphalerite, chalcopyrite, galena, tetrahedrite and arsenopyrite. Gangue minerals include quartz, ferroan carbonate, muscovite and rare barite. The sulphides occur as massive to semi-massive bands in a zone which is overlain by a silica-ferroan carbonate-barite assemblage, interpreted as a possible exhalite (Burgoyne and Giroux, 2011).
ULTRAMAFIC-HOSTED NICKEL-COPPER PGE

Wellgreen (Large)

The large Wellgreen deposit is located 317 km NW of Whitehorse in the southwest Yukon (Figure 36). Mineralization was discovered by prospectors in 1952 and was optioned in the same year to the Yukon Mining Company, a subsidiary of Hudson Bay Mining and Smelting. There was a significant drilling program in 1955 and underground development from 1953 to 1956. Mining and milling began in 1972 but was shortly afterwards suspended due to low metal prices and unexpectedly erratic ore. The latest owner is Wellgreen Platinum Ltd.

The property lies within the Insular Superterrane consisting of the Wrangel and Alexander terranes which were amalgamated at ~320 Ma. The Wellgreen host rock is within the Kluane Ultramafic Belt which lies within Wrangellia and consists of Triassic flood basalts and related intrusive rocks. Below the basalts there are Pennsylvanian and Permian arc volcanics and Permian sediments (Skolai Group) overlain by Middle and Late Triassic basalt and limestone (Nikolai Formation). Sills of the Kluane Ultramafic Belt are common within the Skolai Group and are thought to feed the Nikolai volcanics. The Wellgreen deposit lies along the lower contact of an Upper Triassic sill, locally referred to as the Quill Creek Complex (Carter et al., 2012).

The main component of the Quill Creek Complex is 4.2 km long and 700 m wide consisting of a main intrusion with associated overturned sills. These are crudely layered, variably serpentinized and deformed. Compositions within the Quill Creek Complex include gabbro, clinopyroxenite, peridotite and dunite. The mineralizations are described as the East Zone, West Zone and North Zone of which the East Zone, described here first, has received the most attention. The mineralization occurs at the base of a peridotite body within which there are massive sulphide lenses as well as a skarn zone in calcareous footwall strata: its extent is 1500 m by 700 m wide but of unreported thickness. The East Zone was mined in 1972-73, after which mining operations were suspended. The West Zone has mineralization in gabbro, clinopyroxenite and in the interfingerings of these. Typical ore forming features include nickel-copper sulphides in disseminated, net-textured, semi-massive and massive mineralization (Carter et al., 2012).

Typical mineral assemblages within the Wellgreen ores are pyrrhotite, pentlandite, chalcopyrite, pyrite, magnetite and ilmenite (Photo 35). Less common phases include violarite, sphalerite, chromite, cobaltite, arsenopyrite and 14 others. Specific to the platinum and palladium minerals are arsenides, antimonides, tellurides and bismuthinides of which more than sixteen have been recognized (Carter et al., 2012).
This group of deposits (see Figure 36) includes the skarn-associated epithermal zinc-lead mineralization at the Andrew deposit (veins with zinc and lead), Keglović (skarn associated zinc, copper and lead), Keno Hill (silver, lead, zinc), Logan (silver and zinc) and Quartz Lake (not described).
Andrew (Medium)

The property is located 100 km NE of Faro in the east-central Yukon. Geochemical results acquired by a mining syndicate led to the first staking in 1967. Fieldwork in 1968 led to the staking of the showings that included the Andrew deposit. The deposit is held by Overland Resources Ltd. Resource estimates based on drilling to date were announced in 2008 and continued property work was accomplished through at least 2012.

The property is located within the western Selwyn basin. Host rocks include the Yusezyu and Narchilla formations of the Hyland Group (Neoproterozoic to Early Cambrian), Ordovician to Silurian Road River Formation, and Devonian to Mississippian Earn group. This sequence is intruded by Cretaceous granite, quartz monzonite and granodiorite. The Andrew deposit is underlain by massive quartz arenite interbedded with Narchilla Formation red-green mudstone and lesser limestone. The deposit is described as a polymetallic vein system consisting of multigeneration breccia and veining, extending over an area of 650 m x 250 m (Anonymous, 2014a). The mineralization consists of coarse-grained sphalerite and galena, also as disseminated blebs, veins and massive aggregates. Chalcopyrite occurs in trace quantities.

Keglovic (Medium)

The Keglovic property is located in the south-central Yukon 40 km N of Faro which is 365 km by road from Whitehorse. Exploration from 1965 to 1985 focussed on SEDEX potential (similar to the nearby deposits of the Anvil district): this proved unsuccessful. Occurrence of disseminated, bleb and banded pyrite and pyrrhotite in drill holes was initially ignored. Exploration continued with varying success to 2012.

The property, held by Silver Range Resources Ltd, is located within the Selwyn Basin. Deformation and metamorphism accompanied granitoid intrusion and terrane accretion from the Jurassic to the mid-Cretaceous. Strike-slip faulting, notably the Tintina Fault, was active in the Paleogene. Paleozoic strata on the property are sandwiched between the Anvil Batholith (monzonite, granodiorite) to the southwest and the Teddy Caldera (tuff) to the northeast. The youngest igneous rocks are Paleogene plugs associated with the Ross volcanics.

Intense alteration is associated with the Mount Christie shale (Mississippian) and Tay (Carboniferous, Permian) formations along with sulphide veinlets and disseminated sulphides. Mineralization is associated with the “Keg Main Zone” which occurs in altered and skarn-associated Tay and Mount Christie formations. Coarse grained sphalerite, chalcopyrite and galena occur together with varying amounts of pyrrhotite, pyrite and arsenopyrite collectively making up 1 - 10% of rock volume or 20 - 50% of skarn intervals. Galena abundance increases to the east and into the upper parts of the deposit whereas pyrrhotite and chalcopyrite is enriched downwards and to the west. This implies a more proximal hydrothermal cell at depth and westward (Giroux and Mehlis, 2013).

Keno Hill (Large)

The Keno Hill district (a large resource) is located in the central Yukon 500 km by all-weather road from Whitehorse. Exploration and mining in the Keno Hill district dates back to the early 1900s, beginning with the hunt for gold associated with the Klondike gold rush of 1898. Silver was first discovered in 1901 but mining was delayed to 1913 then deferred due to World War I. New discoveries and a staking rush marked the immediate post-war period: mining resumed in the early 1920s, but was terminated because of World War II. Mining reached a peak in the 1950s but declined in the 1960s due to a lack of new discoveries. Mining, including open pit operations, marked activity in the late 1970s. There were also a number of small scale operations through the 1980s. Production ended in 1989. Remaining assets were purchased by Alexco Resource Corporation in 2006.

Within the mining camp, the bulk of the mineralization occurs in the Basal Quartzite Member of the Keno Hill Quartzite (Mississippian). This unit consists of thinly- to thickly-bedded quartzite and graphitic phyllite. Silver mineralization is hosted by a series of northeast-striking faulted veins with left-lateral and normal displacement. These veins can be
up to 30 m wide. Other faults strike northwest and are seen to offset mineralized veins. The vein mineralogy is complicated by the fact that there have been multiple pulses of hydrothermal activity acting on the system. This has produced vein reactivation and related brecciation. Silver mostly occurs in argentiferous galena, argentiferous tetrahedrite (freibergite) and locally in native silver, polybasite, stephanite and pyrargyrite. Other typical sulphides include galena, sphalerite, pyrite, pyrrhotite, arsenopyrite and chalcopyrite (Figure 37). Zoning has also been suggested with the original belief that the high silver grades only occur at shallow depths, above 120 m. However there are deposits with good silver grades to depths of at least 370 m. Features diagnostic of better or more continuous veins include quartzite or greenstone on a vein wall, veins adjacent to cross-faults, vein splitting, and changes in the dip of the veins (Taylor and Arsenault, 2014).

The Keno Hill district is a polymetallic silver-lead-zinc camp with similarities to Coeur-d’Alene, Idaho, mineralization in the Harz Mountains in Germany, and Pribram in the Czech Republic (Cathro, 2006).

Logan (Medium)

The Logan silver-zinc deposit is located 108 km NW of Watson Lake, Yukon and 38 km N of the Alaska Highway. The property was first staked in 1979 by Regional Resources Ltd, who carried out geochemical and geophysical investigations on the property until 1982. Ownership was jointly controlled with Getty Canadian Minerals Ltd from 1984 but was transferred to Fairfield Minerals Ltd in 1986 who expanded the claim block, added an airstrip (1987) and completed a campaign of trenching and drilling (to 1988). The latest owners are Almaden Minerals (2015).

The zinc and silver occur in a fault-bounded body 1100 m long and 50 - 140 m wide. The deposit is open to a depth of below 200 m. Ore minerals include sphalerite, pyrite, arsenopyrite, chalcopyrite, minor silver sulphosalts and cassiterite. There are also traces of pyrrhotite, covellite, galena, chalcocite, tetrahedrite, stannite, jamesonite, kobellite and native copper. These are concentrated in multiphase quartz- and quartz-ankerite veins, breccia bodies, stockwork and silicified zones. Alteration phases include sericite, biotite and silica (Anonymous, 2014c).
This group includes the Brewery Creek, Golden Revenue (large), Mount Nansen (not described) and Skukum deposits, all located near to, or S of the Tintina Fault in the southwestern Yukon (Figure 36).

**Brewery Creek (Medium)**

The Brewery Creek property lies in the foothills of the Ogilvie Mountains, 55 km E of Dawson City in the northwest Yukon. Claims were first staked by Noranda Exploration in 1987. Northern Tiger then took over the Brewery Creek project in late 2013. The current operator is Golden Predator Corporation.

The Brewery Creek site lies along the northeastern edge of the 15 km wide Tintina fault zone within the western margin of the Selwyn Basin. Strata within the property are composed of deep-water Cambrian to Mississippian strata. These are cut by Cretaceous intrusives of the Tombstone Suite and have suffered deformation in association with transport on the Dawson, Tombstone and Robert Service thrusts. Intrusives on the Brewery Creek property are sill-like bodies of monzonite and quartz monzonite emplaced in graphitic argillite along the contact between the Earn (Devonian-Mississippian) and Road River (Ordovician-Silurian) groups. Gold mineralization is contained within, or adjacent to these intrusive rocks. Alteration associated with the gold includes carbonate-clay, quartz and pyrite-arsenopyrite alteration of monzonite, quartz monzonite and intruded siliciclastic rocks. There are fourteen named deposits. The Kokanee and Golden deposits occur primarily in a quartz monzonite sill and to a lesser extent in siltstone and argillite. Millimetre-scale veinlets contain oxidized pyrite and quartz (Figures 38, 39). Feldspar phenocrysts and the matrix in the host quartz monzonite are altered to clay (Hulse et al., 2014). The Brewery Creek deposit is considered to be an alkalic intrusion-related epithermal deposit.

**Golden Revenue (Large)**

The property is located 200 km NW of Whitehorse. Lode gold was first discovered on Freegold Mountain in 1930, which precipitated a staking
rush through to 1931. Work continued intermittently through the 1950s when interest shifted to porphyry deposits. A soil survey in the 1960s led to the discovery of the Nucleus deposit. Northern Freegold Resources Ltd acquired the property in 2006 and has since conducted extensive surveys.

The Golden Revenue region lies within the Dawson Range, part of a belt of Jurassic to Late Cretaceous plutons that extends to the Alaska border along the boundary between the Stikine and Yukon-Tanana terranes. Components of the Yukon-Tanana terrane include Early Mississippian plutonic rocks and Mississippian and older undifferentiated metasediments and metavolcanic rocks. The property lies within the Yukon-Tanana terrane comprising Paleozoic (or older) quartz-feldspar-mica schist and gneiss, chlorite schist, amphibolite, grey marble and quartzite. There is a penetrative foliation striking northwest and dipping northeast. Granitoid rocks in the area are Jurassic and Cretaceous in age: syenite is found locally. These intrusives are all intruded by Late Cretaceous plugs, sills, dykes and breccia bodies, of felsic to intermediate composition, closely associated with mineralization.

The mineralization is contained between two regional fault systems with a northwesterly trend: the Big Creek fault and the Southern Big Creek Fault. Between these are secondary structures that strike variably west to northwest and northwest to north and northeast. Mineralization is closely associated with west- to northwest-striking structures.

The Revenue Breccia is closely associated with the Revenue deposit. Its matrix is quartz feldspar porphyry. The breccia is typically altered to clay and carbonate and also contains pyrite and copper oxides. Mineralization is contained within the Revenue breccia and in host granodiorite. It is typically seen as porphyry veins, stockwork and disseminated sulphides. Minerals of economic interest include native gold, chalcopyrite and silver with lesser molybdenite and scheelite. Alteration includes potassic (as biotite), sericitic and clay varieties (Armitage et al., 2012).

**Skukum (Medium)**

The Skukum property (owner: New Pacific Metals Corp) is located 80 km S of Whitehorse. There are three deposits on the property: Mt. Skukum discovered in 1982, Skukum Creek (1922) and Goddell Gully (1898). The oldest rocks are assigned to the metamorphosed Nisling and Nasina assemblages (Proterozoic and Paleozoic). Mesozoic plutonic rocks, however, underlie much of the Skukum region. These intrusive Jurassic andesite and siliciclastic sediments and are overlapped by strata of the Whitehorse Trough (Simpson, 2013).

Productive mineralization consists of electrum in quartz-calcite-sericite veins. At Mt. Skukum this includes the Main Cirque Zone (veins and stockwork 200 m long x 5 m wide) emplaced into porphyritic andesite. The Lake Zone (650 m x average 0.6 m wide), consisting of veins, hydrothermal breccia and stockwork, features pyrite, pyrrhotite, sphalerite, galena, rhodochrosite, rhodonite and electrum in a gangue of quartz, calcite and sericite. At Skukum Creek, mineralization occurs in andesite and siliciclastic sediments and is overlapped by strata of the Whitehorse Trough (Simpson, 2013).
This group includes intrusion-related gold deposits of the central Yukon including the Lone Star (not described), Dublin Gulch and Hyland gold deposits (Figure 36).

Dublin Gulch (Medium)

The property is located 85 km by road N of Mayo and 400 km N of Whitehorse. The Dublin Gulch property has a history dating back to 1895 with the onset of placer mining. Exploration of bedrock has been more-or-less continuous since 1970 initially for tungsten, later for gold. Victoria Gold Corp. acquired the property in 2009. The property lies within the north-central part of the Selwyn Basin, which consists of four local units: Neoproterozoic to lower Cambrian Hyland Group, Keno Hill Quartzite and Upper Schist (Devonian-Mississippian), and Lower Schist (Mesozoic). These are associated with three principal thrusts: the Dawson, Tombstone and Robert Service thrusts of which the Robert Service Thrust is closest to the property. It places the Hyland Group over the Keno Hill Quartzite. The thrusting is closely associated with the development of a penetrative cleavage with superimposed east-trending and south-plunging folds.

Phases of intrusion include mid- to late-Cretaceous granitoids: the Selwyn Suite (104 to 98 Ma); the Tombstone Suite (94 to 92 Ma) and the McQueston Suite (64 Ma). These intrusives are closely associated with gold, silver, lead and zinc mineralization. The composition of the Dublin Gulch stock includes granodiorite (oldest and most important), quartz diorite, quartz monzonite and dykes and sills of aplite (Mosher and Tribel, 2011). Mineralization on the Dublin Gulch property is hosted by Hyland Group clastic rocks in close proximity to the Dublin Gulch granodiorite intrusion (93 Ma), part of the Tombstone Suite.

The Eagle zone is located in the Dublin Gulch stock near its western limit with the Hyland Group.
Group. The mineralization consists of sub-parallel quartz veins of grey quartz and potassium feldspar. Vein densities average 3-5/m. Sulphide phases include pyrrhotite, pyrite, arsenopyrite, chalcopyrite, sphalerite, bismuthinite, molybdenite and galena (Figure 41). Alteration assemblages typically include potassium feldspar and sericite-carbonate. Potential mineralization may be located in the Hyland Group near the Dublin Gulch stock (Mosher and Tribel, 2011).

The Eagle Zone on the Dublin Gulch property is considered to be a reduced, intrusion-related gold system. This type occurs as a vein array in the carapace of small plutons where they form bulk-tonnage, low-grade deposits with an association of gold, bismuth, tellurium and tungsten.

Hyland Gold (Small)

The Hyland property is located 74 km NE of Watson Lake in the southeast Yukon. It is located within the southeastern Selwyn Basin, consisting of Neoproterozoic to Devonian shale, deep water limestone, chert and grit. Mineralization in the Main zone consists of semi-massive to massive sulphide (pyrite, lesser arsenopyrite) that at the surface is widely oxidized to limonite, goethite, hematite and manganese oxides. In addition there are vein sets consisting of Type I: pyrite and arsenopyrite; Type II: quartz, pyrite, arsenopyrite, chalcopyrite, bismuthinite, tetrahedrite, native gold and, youngest; Type III: quartz, iron carbonate, pyrite, and titanite. Gold grains, up to 35 microns, occur as inclusions in pyrite or along pyrite grain boundaries (Armitage and Gray, 2012).

OROGENIC GOLD

These are orogenic gold deposits distinguished from those of Archean and Paleoproterozoic ages by their association with Cordilleran deformation in the Jurassic and Cretaceous. The group includes the large Coffee deposit and the White Gold property, both located S of Dawson in the northwest Yukon (Figure 36).

Coffee (Large)

The Coffee deposit is located 130 km S of Dawson and 160 km NW of Carmacks. The property area has a sporadic history of placer mining up to 1982. Recent hard-rock gold exploration began with the discovery in 1981 of arsenic anomalies in Coffee Creek. Gold anomalies in soils were uncovered in 1999 over an area of 400 m x 900 m. Kaminak Gold Corp. optioned the property in 2009 and expanded a program of exploration and drilling down to the present (2013).

The Coffee property lies within the Yukon-Tanana terrane (YTT). The YTT consists of pre-Devonian schist, local amphibolite, greenstone and ultramafic rocks. An intrusive arc (Klondike arc) developed on the Yukon-Tanana terrane in the Late Permian, leading to collision with the Laurentian margin in the Jurassic. Ongoing shortening produced folding and thrusting of greenstone and ultramafic slivers (Slide Mountain terrane) within the YTT. These events are dated to the Early Jurassic. Extensional emplacement of local granite is of mid-Cretaceous age (~110 - 90 Ma).
The Coffee property is underlain by Paleozoic metasediments, Cretaceous Coffee Creek granite and a portion of the Dawson Range batholith. Deformation fabrics include metamorphic foliation, shearing (Jurassic) and brittle fracturing and faulting. These latter structures host gold mineralization and are identified as splays of the nearby Big Creek fault which displaces the Cretaceous granite and are therefore younger. Fault-associated structures include crackle breccia and stockwork fracture systems. In total nineteen gold zones have been discovered (Sim and Kappes, 2014).

Hydrothermal breccia is the dominant host in the Supremo zone (Figure 42) but lower-grade mineralization is also found in hydrothermally altered gneiss and in altered dykes. Minerals found in association with micron-scale and “invisible” gold include pyrite, micron-sized barite, an iron-barium arsenate, a phosphorus phase, monazite and zircon (Mackenzie et al., 2015). Supergene alteration, expressed as iron oxides, is present to depths exceeding 200 m. It has been suggested that the Coffee deposit is of the epizonal high-sulfidation hydrothermal orogenic gold type.

White Gold (Medium)

The White Gold property is located 95 km S of Dawson City and 350 km NW of Whitehorse at the confluence of the Stewart, White and Yukon rivers. In spite of a long history of placer mining, hard rock exploration did not begin in earnest until 2007 when the property was optioned to Underworld Resources Inc. It is now held by Kinross Gold.

White Gold occurs in greenschist and lower amphibolite facies, metamorphic rocks of the Yukon-Tanana terrane which was accreted to ancestral North America in the Jurassic. Thrusting marked the Jurassic and plutonism the Jurassic and Cretaceous. The deposit is now located between the Tintina and Farewell faults in the central Yukon.

Three geological zones are exposed on the property: quartzite in the west; overlying strongly foliated and lineated metavolcanic rocks (amphibolite gneiss) in the central region; and an eastern belt of metasediments (quartz-rich unit and schist) intruded by pyroxenite and serpentinite. Two ore zones have been delineated at White Gold: Golden Saddle and Arc. The Golden Saddle mineralization is located within felsic orthogneiss, amphibolite and ultramafic
rocks. Fault zones and breccia have important associations with the mineralization. Gold is specifically associated with vein- and disseminated pyrite, lode and stockwork quartz veins, quartz vein breccias, silicification, and limonite (Figure 43). Lode veins and breccias also carry minor molybdenite, galena and chalcopyrite. Gold occurs as blebs up to 15 microns close to or within pyrite (Weiershauser et al., 2010).

Gold in the Arc zone is mostly present in metasedimentary rocks, specifically in quartzite and biotite schist, but it is also found in felsic to intermediate dykes. Alteration is present as silicification and graphite. Mineralization occurs in veinlets as micron-scale gold in arsenopyrite along with pyrite, pyrrhotite, graphite and minor sphalerite (Weiershauser et al., 2010).

**MANTO GOLD**

The Manto gold deposits are located at Ketza River (not described), close to, but S of the Tintina Fault in the southern Yukon, and at Rakla-Rau in the Mackenzie Mountains (Figure 36).

**Rackla - Rau (Medium)**

The Rau property is located in the northern Yukon 100 km NE of Mayo. Exploration began in 1922 with the discovery of mineralized float within the property area. More silver-bearing galena float was discovered by Geological Survey of Canada staff in 1924. However, the source was not located. The first “in situ” discoveries were by an independent prospector in 1923. Claims were periodically staked and dropped up to 1974. The history of ATAC Resources Limited in the property and of accelerated exploration dates to 2006.

The Rau property is located within a zone of thrust imbrication that includes rocks of the Mackenzie Platform (shallow water carbonates and clastics) and Selwyn Basin (Neoproterozoic to Paleozoic basinal rocks). These are intruded by Late Cretaceous felsic plutons of the Tombstone Suite and by felsic intrusions of the McQueston Suite (65 Ma). There is, within the property, a region of dykes and sills considered to be the upper portion of a largely unroofed granite body referred to as the “Rackla Pluton”. Ar/Ar ages are in the range of 59.1-62.4 Ma. This area is a locus for skarn development (Strosheim et al., 2011). Gold mineralization in the Tiger zone is developed in a carbonate replacement body 700 m x up to 200 m wide x up to 96 m thick in a high strain zone. The mineralization occurs in folded carbonate beds interleaved with mafic flows and volcaniclastic units. Gold occurs primarily in a sulphide facies, which includes banded pyrite, arsenopyrite and pyrrhotite with minor bismuthinite, sphalerite and scheelite (Strosheim et al., 2011). Carbonate textures indicate extensive karst dissolution and phreatic zone precipitation. Void fill is commonly pyrite, spar calcite and quartz. Alteration consists of talc and sericite, which is limited to volcanic horizons. The Tiger zone mineralization shares features with both the low-temperature Carlin-type gold and others of a higher-temperature genesis.
PORPHYRY COPPER AND MOLYBDENUM

Classic porphyry deposits of the Yukon (Figure 44) variously comprising chalcopyrite, molybdenite, gold and tungsten minerals include Cash (medium: copper, molybdenum and gold), Casino (very large: copper and molybdenum), Logtung (large: tungsten and molybdenum) and Red Mountain (large: molybdenum).

Figure 44. Simplified geology of Yukon and western Northwest Territories also showing deposits of tungsten and copper skarns, porphyrys and iron oxide copper gold.
Cash (Medium)

The Cash property is located 452 km NE of the seaport of Skagway, Alaska in the vicinity of Carmacks in a belt of porphyry deposits extending 100 km from Casino in the northwest to Antoniuk in the southeast. Copper, molybdenum and gold-porphyry style mineralization underlies an area of 3050 m x 900 m associated with copper-molybdenum soil geochemical anomalies and with mid-Cretaceous Mt. Nansen Group feldspar porphyry dykes and plugs. Older syenite and quartz monzonite stocks, and schist and gneiss of Paleozoic age are also present. The mineralization consists of disseminated and fracture-filling chalcopyrite and pyrite with minor coarse-grained molybdenite in quartz veinlets. Zoning indicates high molybdenum values in the northeast and higher gold in the southwest. Systematic drilling in 1975-1977 defined a northwestern limit to the mineralization. The ore zone remains open on the other three sides and at depth. The deposit is expressed at the surface by a 900 m x 2500 m copper and molybdenum anomaly, which is fringed by anomalies of lead, zinc, silver and gold. Soil sampling within apparent fault zones has produced additional soil geochemical anomalies for gold and arsenic (Anonymous, 2014g).

Casino (Very large)

The very large Casino deposit is located in the Dawson Range Mountains 300 km NW of Whitehorse in the west central Yukon. Placer claims were first staked in 1911. A 1917 report by D.D. Cairns of the Geological Survey of Canada suggested that gold and tungsten might be found in a nearby intrusive complex, on what would become the location of the Casino deposit. Silver-lead-zinc veins were discovered in 1936 and were the focus of exploration up to 1967. Silver-rich veins were mined periodically from 1963 to 1980. Porphyry potential was investigated from 1967 onwards with a variety of owners and joint-venture partners. The current owner is Casino Mining Corporation.

Rocks of the Dawson Range are represented by the Devonian-Mississippian Wolverine Creek metamorphic suite consisting of metasedimentary quartz-feldspar-mica schist, gneiss, quartzite and meta-igneous biotite-hornblende-feldspar gneiss, othogneiss and amphibolites.
chalcanthite, malachite, brocanthite and others. The supergene sulphide zone is also notably enriched in copper and features digenite, chalcocite, minor covellite, bornite and copper-bearing goethite. Below the weathered zone, mineralization is typically found in stockwork veins and breccias. Hypogene mineralization in the potassic zone consists of finely disseminated pyrite, chalcopyrite, molybdenite and minor sphalerite and bornite. Gold, copper, molybdenite and tungsten are present in higher grades in the phyllic zone (Figure 45; Huss et al., 2013).

**Logtung (Large)**

The large Logtung deposit is located 260 km SE of Whitehorse, Yukon and 165 km W of Watson Lake near the BC-Yukon border (but entirely within the Yukon). The property lies 130 km SW of the Tintina Fault in the Yukon-Tanana terrane. The country rocks consist of Paleozoic and Triassic clastic and carbonate sedimentary rocks. These are intruded by two plutonic suites of Early Jurassic and mid-Cretaceous ages. The Early Jurassic magmatic rocks feature en echelon plutons ranging from ultramafic to granodioritic in composition with a northwesterly trend. The Cretaceous plutonic suite (115 - 97 Ma) also trends northwest and has a compositional range from quartz monzonite to monzogranite. The suite includes batholiths, plutons and hypabyssal dykes and is the host of the Logtung deposit (Molavi et al., 2011).

The Logtung porphyry tungsten-molybdenum deposit is characterized by a quartz-vein stockwork and a sheeted-vein set centred on a quartz-feldspar intrusion complex. This is a branch of a mid-Cretaceous quartz monzonite stock, a satellite of the Seagull batholith, and one of several intrusions that are enriched in tungsten, molybdenum and fluorine. Where skarn is present the ore minerals are associated with veins and with open fractures. For this reason Logtung is considered to be a porphyry deposit rather than skarn mineralization (Noble et al, 1984).

The property is, nevertheless, underlain by skarn and hornfelsed metasedimentary rocks. The skarns include green quartz-diopside skarn and reddish-brown garnet skarn interbedded with grey to black hornfels. Intrusive rocks include Jurassic diorite dykes, a mid-Cretaceous quartz monzonite stock and related quartz-feldspar porphyry dykes. The mineralized zone is 2.5 km x 1.0 km and extends along the northern and western margins of the quartz monzonite stock. Mineralization also extends into the stock and is closely associated with porphyry dykes. Minerals of economic significance include scheelite, molybdenite and molybdoscheelite. There is also a sulphide facies associated with northeast-striking sheeted quartz-pyrite veins that includes beryl, fluorite, bismuthinite, chalcopyrite, sphalerite and pyrrhotite (Molavi et al., 2011). The Logtung deposit is classified as a porphyry of tungsten-molybdenum type. These are large-tonnage, low-grade hydrothermal deposits.

**Red Mountain (Large)**

The Red Mountain property is located 130 km SSW of Ross River in the south central Yukon. The area was first staked in 1967. Ground surveys and drilling commenced in 1969 with operators including Hudson Bay Oil and Gas Co.Ltd. and Amoco Canada Petroleum Co. Ltd. An application to conduct underground exploration was made by Tintina Mines Ltd. in 2005.

Paleozoic argillite of the Yukon-Tanana terrane is intruded by a quartz monzonite porphyry stock in which molybdenite occurs in a quartz stockwork. Mineralization extends also partly into the surrounding argillite. Geochemical anomalies over a pyritic gossan include molybdenum, copper, silver and tungsten. The porphyry stock has concentric alteration haloes and is cut by a post-mineralization quartz diorite stock. The mineralization underlies an area of 1500 m x 425 m and extends to a depth of >1125 m depth (Anonymous, 2014d).
COPPER AND TUNGSTEN SKARNS

Skarn mineralization has developed adjacent to Cretaceous intrusive rocks most notably along the Yukon-Northwest Territories border (Figure 44). Here is found the Cantung tungsten mine and the economically significant deposits that include the large Mactung (inside the Yukon) and the Lened deposit (inside the Northwest Territories; not described). Also described is the Risby tungsten property west of Ross River, the Sa Dene Hes deposit near Watson Lake, and the copper skarn deposits at Whitehorse.

Cantung (Medium)

The Cantung deposit is located in the Nahanni region of the Northwest Territories, 300 km by all-weather road NE of Watson Lake, Yukon and close to the Yukon border. The Cantung mine lies close to the boundary between the Mackenzie Platform (to the east) and the Selwyn basin (to the west). Stratigraphic units in the mine area include early Cambrian limestone (“Swiss Cheese Limestone”) overlain by an early Cambrian clean blue grey limestone (the Ore Limestone) succeeded by argillite, quartzite, dolostone (all Early Cambrian) and above this wavy-banded limestone of the Middle and late Cambrian Rabbitkettle Formation (Fitzpatrick and Bakker, 2011).

The dominant structure is the Flat River Syncline which is ca. 5 km wide and trends northwest. Intrusives of Cretaceous age are post-tectonic and consist of medium-grained biotite quartz monzonite. There are two of these at surface in the immediate vicinity of the mine and a third at depth associated with dykes that cut overlying strata. These are aplite, leucocratic- or tourmaline-bearing granite and porphyry dykes (Fitzpatrick and Bakker, 2011, Rasmussen et al., 2011).

Parts of the deposit, Open Pit and E Zone, lie on the flat-lying upper limb of a local recumbent anticline and on the lower limb, respectively. The mineralization is associated with a calcsilicate skarn in the Ore Limestone within which scheelite is the sole ore mineral (Figure 46). However, chalcopyrite was formerly also acquired during mining. Gangue phases in the Open Pit deposit include pyrrhotite, diopside, garnet and actinolite. The E Zone deposit consists of massive to semi-massive pyrrhotite, pyroxene, garnet,
actinolite and biotite. Accessory phases include apatite, epidote and tourmaline (Fitzpatrick and Bakker, 2011).

**Mactung (Large)**

The Mactung property (Narcisco et al., 2009; Selby et al., 2003), held by North American Tungsten Corporation Ltd, straddles the Yukon - Northwest Territories border NW of Cantung. It is located in the eastern Selwyn Basin, a region of off-shelf, deep-water sediments that persisted from the Neoproterozoic to the Middle Devonian. Typical deep-water strata include shale, chert and basin-facies limestone. These grade northeastwards into shelf facies carbonates. Deformation was largely of Jurassic and Early Cretaceous age, featuring thrust faulting and open to tight folds. Granite magmatism was prominent in Early to Late Cretaceous time. Five intrusive suites are recognized and range form 97.5 - 92 Ma (Selby et al., 2003).

The Mactung deposit is the most northerly of a 200 km long northwesterly trending belt of tungsten-copper skarns that also includes the Cantung deposit, 160 km to the southeast. A typical feature of these deposits is the location of skarns above the altered apex of Late Cretaceous quartz monzonite plutons. This feature is found in the Mactung deposit, which is closely associated with the Cirque Lake stock.

The mineralization is developed in scheelite skarns within an interbedded succession of Cambrian to Silurian limestones, shales and siltstones near the south contact of the Cirque Lake stock. There are two skarn zones, separated by 100 m of hornfelsed shale and siltstone variously altered to muscovite, biotite and graphite. Cutting these there are numerous veinlets containing pyrite, pyrrhotite, scheelite and molybdenite (Narcisco et al., 2009). The entire section of nine mappable units, four of which contain tungsten ores, forms a recumbent fold with a gently plunging axis. The ore body is also cut by numerous steeply dipping faults with displacements of up to 45 m and consisting of gouge, breccia and quartz and calcite pore-filling.

The mineralization consists of a pyroxene-pyrrhotite-scheelite skarn (Figure 47) and, less commonly, a garnet-pyroxene-scheelite skarn. Disseminated pyrite occurs in some phyllite units, and galena and sphalerite in quartz veinlets. Wolframite and chalcopyrite are rare.

![Figure 47. Mactung: scheelite-bearing massive pyrrhotite (Yukon Geological Survey).](Image)
Pyrrhotite and scheelite occur in veins, fracture fillings and disseminations in the upper skarn zone. The Mactung deposit is classified as a contact-metasomatic skarn (with hydrothermal fluids originating in a nearby stock).

Risby (Medium)
The Risby property is located 55 km W of Ross River. Lower Paleozoic rocks are intruded by Cretaceous biotite-quartz monzonite which has produced scheelite skarns containing tungsten with lesser copper and molybdenum. The skarn zones are parallel to the intrusive contact and are identified as lower and upper zones. The known length of the deposit is at least 750 m (as of 2009) (Desautels, 2009). The property was worked from 1968 to 1982 by the Caltor Syndicate and Hudson Bay Exploration and Development Co. Ltd. This included mapping, trenching, stream sediment sampling, ground geophysics and drilling. The drilling was subsequently continued by Playfair Mining Ltd who also obtained a compliant resource estimate.

Sa Dene Hes (Medium)
The Sa Dene Hes property, 50 km N of Watson Lake, consists of four significant occurrences. Mineralization was first discovered in the area in 1962. The Main zone was delineated by surface drilling and other activities by Cima Resources from 1979 to 1981. Underground development by the Mt. Hundere Joint Venture occurred in 1990. The mine opened in 1991 but was soon closed as a result of low metal prices.

The host rocks include pelitic phyllite and limestone. The limestone occurs as discontinuous units up to 100 m thick, each traceable for hundreds of metres and gradationally laterally with phyllite. Archeocyathids indicate an early Cambrian age. The phyllites vary from brown non-calcareous phyllite to grey calcareous phyllite and carbonaceous phyllite. There are three suites of igneous rocks including mafic to intermediate chloritic intrusive, a fine-grained unfoliated intermediate intrusive including skarn, and quartz porphyry occurring as dykes (Anonymous, 2014e).

Skarn mineralization, also associated with hornfelsing, occurs along contacts between limestone and phyllite. The skarn assemblage includes actinolite, hedenbergite, diopside, grossular, andradite, chlorite, calcite and quartz. Wollastonite, fluorite and amethyst have also been reported. The sulphide mineralization features medium to coarse sphalerite and galena occurring as disseminations in skarn. Peripheral to this there are local magnetite skarns (Anonymous, 2014e).

Whitehorse copper (Medium)
The Whitehorse copper belt is located SW and NW of downtown Whitehorse and, at the closest, 5.6 km from the town. The copper belt extends in a northwesterly direction for 32 km along the western margin of the Whitehorse Batholith.

The Whitehorse copper belt consists primarily of skarn deposits found within Late Triassic Lewes River Group but also within mid-Cretaceous granodiorite. Mineralization tends to occur within embayments of the adjacent batholith but also extends up to 150 m from the intrusive contact. Ore deposits range from lenticular or tabular to irregular but are normally conformable with bedding. The best ore occurs on contacts between limestone and quartzite. Two types of skarn are distinguished: magnetite skarn, and; silicate skarn. Magnetite skarn features bornite, chalcopyrite, magnetite, serpentine, specularite, talc, chlorite and minor pyrrhotite and pyrite. The mineral assemblage in silicate skarn includes bornite, chalcopyrite, garnet, diopside, wollastonite, tremolite, epidote, chlorite, calcite and quartz (Anonymous, 2014h). There is also a relationship between skarn mineralogy and protolith composition. Limestone protoliths form skarns with andraditic garnet, iron-rich pyroxene, wollastonite and vesuvianite. Dolomite protoliths form skarns with diopside, forsteritic olivine and andradite with retrograde phlogopite, brucite, serpentine and talc (Dawson and Kirkham, 1995). Sulphides tend to be associated with the retrograde phases: chalcopyrite and pyrite with retrograde actinolite and chlorite; bornite and chalcocite with epidote. Other metal-bearing phases include chrysocolla, native copper, chalcocite and minor molybdenite, scheelite, tetrahedrite and valerite (a micaeous iron copper hydroxy-sulphide). There are, in addition, indications of elevated contents of gallium and silver in association with copper.
Although hundreds of kimberlite pipes have been discovered in the Archean Slave craton and in Archean portions of the Rae and Hearne cratons, most of these are isolated and small or unproductive. One of the key aspects to economic significance is sufficient tonnage to justify a
mine life of 10 - 20 years. The majority of the kimberlite pipes in northern Canada are small and of low tonnage, hence there has been focus on multiple diamondiferous pipes in close proximity (Kjarsgaard, 2007). Producing mines include Ekati (1998), Diavik (2001), Snap Lake (2008) and Gahcho Kué (2016) in the Northwest Territories part of the Slave craton (Figure 48), as well as Jericho (past producer). Kimberlite fields that have some possibility of becoming mineable include Chidliak on southern Baffin Island (Figure 11) and Qilalugaq on southern Melville Peninsula near Repulse Bay (Figure 2).

Ekati

Ekati is located 300 km NE of Yellowknife in the central Slave craton. It includes the Ekati diamond mine (Figures 49, 50), currently held by Dominion Diamond Corporation, >150 kimberlites and four producing pipes. Most of the kimberlites are small, steep-sided diatreme-like bodies, relatively xenolith-poor and commonly consisting of bedded volcaniclastic materials.

The Slave craton in the Ekati area consists of Archean supracrustal rocks (Yellowknife Supergroup), syn- and post-tectonic granitoid plutons (2.63-2.58 Ga) and five dyke swarms (2.23-1.27 Ga). Xenoliths in the kimberlites indicate the former existence of Late Cretaceous and Paleogene strata (ca. 105 - 45 Ma) buried to a maximum depth of ~1100 m (Nowicki et al., 2004).

The kimberlites have been preferentially emplaced into Archean plutonic rocks, at lineament intersections, along dykes, and at dyke intersections. Thirty kimberlites have ages of 75-45 Ma. Most are steep-sided and inward-tapering cones but there is a spectrum of shapes (e.g. outward-dipping, elongate, irregular). Normal sizes range from 0.1 ha - 5.0 ha and up to a maximum of 17 ha. Panda, for example, has an area of 3 ha at the surface; Koala North: 0.4 ha; Koala: 4.5 ha and; Fox: 17 ha. Although most are single pipes, the Misery complex consists of eight distinct bodies (Nowicki et al., 2004).

Volcaniclastic kimberlite (VK) is a common diatreme fill, with compositional variability based on olivine abundance, grain size, proportion of matrix, scale of bedding, quantity and nature of xenoliths and the character of cognate clasts and lapilli. Types of volcaniclastic kimberlite include resedimented (RVK), tuffisitic (TK), mud-rich (m-) and olivine rich (o-). Magmatic kimberlite (MK) is less common. The proportion of wall rock xenoliths does not generally exceed 5% (Nowicki et al., 2004).
Mud-rich resedimented volcaniclastic kimberlite (mRVK) is ubiquitously common in the Ekati pipes. These are interpreted to have formed by mass flow (debris or mud-flow) processes. The olivine-rich kimberlites (oVK) have >40% olivine. The matrix consists of disaggregated mud, kimberlitic ash and serpentine. Unaltered wood is also common. Olivine-rich kimberlite deposits are interpreted to have formed as crater-rim deposits by cool emplacement processes.

Olivine-poor sedimentary rocks, generally a minor component in the upper part of many pipes, consist of siltstone, mudstone and sandstone in addition to kimberlitic components such as phlogopite, olivine or mantle xenoliths. These have been collectively identified as crater-fill epiclastic sediments.

Primary (juvenile) VK contains 20 to 50% olivine and lacks bedding but includes abundant lapilli. Lapilli consist of olivine set in a carbonate-rich matrix along with serpentine, opaque phases and perovskite. The magma matrix is mostly serpentine with minor mud or ash, and carbonate: it is a dominant component in the deeper part of various pipes, and specifically in Panda, Koala and Lynx. An unworked, non-resedimented, volcaniclastic origin is indicated.

Tuffisitic kimberlite (TK) occurs in the Fox and Misery pipes. It is a fragmented and clay-altered material with a high concentration (>40%) of comminuted and xenolithic granodiorite, serpentinized olivine (15-30%) and juvenile lapilli set in a matrix of serpentine and clinopyroxene.

Magmatic kimberlite (MK) occurs in both large bodies and in dykes. Minerals present include macrocrystic olivine, macrocrystic phlogopite, xenocrystic garnet, chrome diopside and rare ilmenite. The groundmass includes fine opaque magnesian ulvöspinel, perovskite and monticellite. The matrix of the MK in dykes includes monticellite, phlogopite, carbonate and serpentine. Pipes dominated by MK include Leslie, Grizzly, Pigeon, Arnie and Mark.

Syn-depositional winnowing and mass-flow-related sorting may contribute to improvement of grade. Diamond contents are inversely proportional to diluting materials such as mud and ash. Diamond abundances are closely correlated with the contents of macrocrystic olivine and other mantle-derived materials including eclogitic garnet, chromite and chrome diopside. These minerals have also proved to be useful as diamond indicators during till-sampling campaigns.
Diavik

The Diavik mine property, currently held by Dominion Diamond Corporation, is located 295 km NE of Yellowknife in the central Slave craton. Major features of this craton include a 4.05–3.6 Ga orogenic nucleus, felsic granites and gneisses (3.2–2.8 Ga), and pre- to post-orogenic felsic plutons (2.69–2.60 Ga). There are also dyke swarms of various ages (i.e. 2.3–2.0 Ga; 1.27 Ga). The Diavik property includes four diamondiferous kimberlite pipes located on the bed of eastern Lac de Gras. Initial staking of the property in 1991 preceded the exploration and discovery phase. Exploration has involved a combination of heavy mineral sampling, airborne magnetics and EM followed by drilling of favourable targets. These methods have resulted in the identification of 52 kimberlite pipes. Four, identified to be commercially significant, were discovered in 1994 and 1995: A21, A154 North, A154 South and A418. Kimberlite Rb-Sr ages on mica yield an age range of 56.0 to 54.8 Ma (Graham et al., 1998). Local host rocks include 2.61 to 2.58 Ga-old granite, granodiorite, tonalite and Yellowknife Supergroup greywacke-mudstone. In addition, there are three diabase dyke swarms dated at 2.23, 2.02 and 1.27 Ga (Buchan et al., 2009). Cover rocks at the time of emplacement included poorly lithified Cretaceous (Albian and younger) to Paleocene sand, silt and muds. The four commercial kimberlites are steeply- to vertically-sided pipes that typically narrow downwards to varying degrees, but also flare outwards with depth. Their surface areas are mostly < 2 ha. The facies found include pyroclastic kimberlite, resedimented volcanioclastic kimberlite, lesser hypabyssal kimberlite, and xenolithic sediments (containing clasts of country-rock granite and Cretaceous-Paleogene mudrock). The volcanioclastic kimberlite contains a large proportion of resedimented tephra, and mud or clasts of xenolithic material. Common components of the pyroclastic and resedimented volcanioclastic kimberlite include fragmented mineral crystals, lithic fragments, plant and vertebrate fossils set in a matrix of serpentinite, silt, clay and carbonate (Bryan and Bonner, 2003; Graham et al., 1998).

Features of the pyroclastic kimberlite include: 1) massive to graded beds; 2) lack of evidence of clast reworking; 3) bombs and blocks of host rock material (granite, mudstone, etc.); and 4) particulate mud that is not waterlain. Resedimented volcanioclastic kimberlite has commonly incorporated sediments of Cretaceous and Paleogene ages. In general there is often a lateral gradation in these rocks from kimberlite to kimberlitic sediment to (non-kimberlitic) sediment. At Diavik the volcanioclastic kimberlite contains a large proportion of tephra, xenolith clasts and beds that originate as debris flows. Clasts up to more than 1 m across include kimberlite, granite and mud, along with sand-grade olivine and other minerals. Beds showing alternating mud and sandy olivine are a common feature.

Jericho

The Jericho kimberlite is located 400 km northeast of Yellowknife in the Nunavut portion of the Slave craton, and 150 km NW of the Lac de Gras (Ekati) kimberlite field. Tehera Diamond Corp operated the property as a diamond mine from 2006 to 2008. Shear Minerals Ltd purchased the property in 2010 and briefly reopened the mine in 2012. A Middle Jurassic age is indicated (173.1±1.3 Ma) (Hayman & Cas, 2011). There are fifteen pipes and dykes in the Jericho cluster. Jericho intrudes granitoids of the Contwoyo batholith (2.59–2.58 Ga). Cover rocks at the time of emplacement included Middle Devonian fossiliferous and unfossiliferous limestone with minor shale and sandstone, now exclusively preserved as xenoliths in the kimberlite (Kopylova and Hayman, 2008).
Extensive drilling and underground development shows the Jericho complex to consist of three distinct lobes bounded on the east by a planar dyke of hypabyssal kimberlite. These lobes are independent cone shaped bodies that taper downward. 80% of the central lobe is high-grade, diamond-bearing kimberlite. Kimberlite compositions include hypabyssal kimberlite in early and late stage dyke sets, light and dark varieties of massive pyroclastic kimberlite (MPK1, MPK2), and weakly bedded pyroclastic kimberlite.

Early-stage hypabyssal kimberlite consists of 20 to 30% olivine macrocrysts in a matrix of olivine, serpenitized monticellite, spinel, perovskite, ilmenite, phlogopite and apatite. Late-stage hypabyssal kimberlite has 10-22% olivine phenocrysts in a matrix of carbonate, serpentine, apatite, phlogopite, spinel, ilmenite and perovskite. MPK1 grades to kimberlite breccia with clasts of limestone, granite, diabase and minor sandstone and mudstone. In general, this material is a carbonate-serpentine kimberlite with olivine macrocrysts, xenocrysts, xenoliths, and lapilli in a matrix of serpentine and carbonate. A regional variant is serpentine kimberlite which is distinguished by the absence of carbonate and spinel. In MPK2, 20 - 70% of the olivine is serpentinized and is commonly accompanied by hematite. The matrix is secondary chlorite, carbonate, spinel and phlogopite plates. Juvenile kimberlite clasts contain carbonate, olivine euhedra, monticellite, spinel, and melilite. Weakly bedded pyroclastic kimberlite consists of three components: 1) grains of olivine, garnet and ilmenite; 2) juvenile lapilli; and 3) a serpentinitic matrix. The lapilli consist of serpentinized olivine, carbonate and ilmenite in a groundmass of spinel, perovskite, calcite and serpentine.

Snap Lake

The Snap Lake property lies 220 km NE of Yellowknife in the northeastern Slave Craton of Northwest Territories. It is underlain by the 2.7 Ga Camsell Lake greenstone belt (Yellowknife Supergroup) volcanic rocks and metasediments that are intruded by gabbro and leucogabbro sills, granite- granodioritic stocks and dykes, and diabase dykes (2.21 - 1.27 Ga).

A joint venture group that included Winspear Resources Ltd and Aber Resources Ltd was assembled in 1993 and undertook airborne surveys and mineral train studies. This led to drilling and the discovery of the CL25 pipe and of the CL174 pipe in 1995. Kimberlitic boulders with diamonds were discovered along the shore of Snap Lake in 1996. Drilling in 1997 determined that the Snap Lake kimberlite was a north-pointing sheet with a thickness of 2.5 m and dipping 150° E. Drilling in 1998 showed that the sheet extends down dip for at least 1800 m. Since then there has been surface and underground testing including bulk sampling. De Beers purchased Winspear Resources Ltd in 2000. The following year De Beers Canada purchased the Aber Resources Ltd interest in Snap Lake. The mine was opened in 2008.

The Snap Lake kimberlite (523 Ma) is a gently dipping sheet that is 2.8 m thick on average (although highly variable in thickness) which is traceable along strike and across strike for 4 km. Within the sheet complex there is a kimberlite pipe (or “blow”), 75 - 100 m in diameter and consisting of granite clast breccia near the surface, grading downwards to lapilli-bearing pyroclastic kimberlite (Kopylova et al., 2010). Locally the kimberlite sheet bifurcates to form numerous stringers or “horsetails”. Faults are also known in the vicinity of the dyke, but these are Proterozoic features and no dyke offset is indicated. Locally it appears that the dyke thickens into the fault zone and has thus exploited this line of weakness.

Groundmass components include phlogopite (20%), monticellite, spinel, apatite, dolomite and serpentine. Opaque phases include chrome spinel and magnetite. Macrocrysts include mostly serpentinized olivine (up to 10 cm), and rare garnet and chromite (Kirkley et al., 2003; Kopylova et al., 2010). Ten percent of the Snap Lake dyke consists of kimberlite breccia with granite and metavolcanic clasts in a phlogopite-rich matrix. A separate, rare component is aphanitic kimberlite which is devoid of macrocrysts and composed largely of serpentine, calcite and opaque phases. A volatile-rich origin, similar to that of carbonatite, has been proposed (Kirkley et al., 2003).
Gahcho Kué

The Gahcho Kué property, held by De Beers Canada Exploration and Mountain Provinces Diamonds Inc., is located 300 km E of Yellowknife and 60 km N of the East Arm of Great Slave Lake. Four significant, named kimberlite bodies (Hearne, 5034, Tuzo and Tesla) are located on the property.

The Gaucho Kue kimberlite cluster is located in the southeastern part of the Archean Slave Craton. The kimberlites are early Cambrian age (542-534 Ma) (Heaman et al., 2003 or 2004). Host rocks on the Gaucho Kué property include 2.63 – 2.58 Ga granite, granitic gneiss, minor granodiorite and diorite that have been metamorphosed and retrograded to greenschist facies and intruded by dykes inferred to be part of the Malley swarm.

Kimberlite facies include tuffisitic kimberlite, transitional tuffisitic kimberlite, transitional hypabyssal kimberlite and hypabyssal kimberlite. The tuffisitic kimberlite is matrix-supported breccias containing 30 - 95% often unaltered granitoid clasts and, less commonly, clasts of diabase, gneiss and volcanic rocks. Other features include serpentinized olivines, pelletal lapilli, matrix serpentine and clay. Transitional tuffisitic kimberlite is darker and country rock xenoliths are less common. Transitional hypabyssal kimberlite is dark and competent. Groundmass minerals include phlogopite, spinel, carbonate, serpentine and perovskite. Hypabyssal kimberlite is fresh, competent and black to dark green. There are two phases of fresh (unaltered) olivine: medium grained anhedral and smaller subhedral and euhedral. The matrix is well crystallized and includes monticellite, phlogopite, spinel, carbonate, serpentine and perovskite. Mantle xenocrysts include garnet and clinoxyroxene. There are also mantle xenoliths of garnet lherzolite and eclogite (Johnson et al., 2014).

The Hearne pipe consists of two bodies underlying a total area of 1.5 ha. Fill consists primarily of tuffisitic kimberlite. The 5034 pipe has five discrete lobes three of which coalesce at depth. There are also two small satellite intrusions. The total area of 5034 is 2.1 ha. Compositions in 5034 include hypabyssal kimberlite at depth, transitional kimberlites above this and tuffisitic kimberlite at the highest levels.

The surface area of the Tuzo pipe is 1.2 ha. Five kimberlite facies have been logged at Tuzo, ranging from tuffisitic in the immediate subsurface to hypabyssal at considerable depth. Other facies include country rock breccia with kimberlite, country rock breccia and an epiclastic unit.

Chidliak

The Chidliak property, held by Peregrine Diamonds Ltd, consists of 60 prospecting permits located on Hall Peninsula (southern Baffin Island) approximately 150 km NE of Iqaluit (the capital of Nunavut). Bedrock on the Hall Peninsula includes Archean basement of the Meta Incognita microcontinent (2920 to 2797 Ma), Paleoproterozoic Lake Harbour Group, and the eastern margin of the Paleoproterozoic Cumberland batholith. Phases of deformation of the microcontinent occurred at 277 Ma and again at 1844 - 1736 Ma. The bulk of the Chidliak property consists of Ramsay River orthogneiss (3019 to 2784 Ma). Northerly-trending belts of Lake Harbour Group are also present. These consist of psammites, quartzite, semi-pelite, pelite, minor marble, calcilicate and leucogranite. Intermediate, mafic and ultramafic igneous rocks include leucodiorite, tonalite, peridotite, pyroxenite and dunite; all metamorphosed to varying degrees (Pell, 2008).

Kimberlites on the Chidliak property include sheets and small pipe-like bodies. The sheet-like bodies consist of hypabyssal (coherent) kimberlite with basement xenoliths. The kimberlite pipes contain basement xenoliths and fragments of carbonate and clastic rock of Late Ordovician to Early Silurian age. These rocks are not encountered at the surface so it is presumed that the Ordovician-Silurian was a former cover on Hall Peninsula. The kimberlites provide a U-Pb age-range on perovskite of 156.7 – 138.9 Ma (Kimmeridgian to Valanginian; Heaman et al., 2015). Four kimberlite facies have been distinguished - volcanic kimberlite (VK), hypabyssal kimberlite (HK), coherent kimberlite (CK) and pyroclastic kimberlite (PK). All of these contain mantle xenoliths. Some, however, contain clasts of gneissic basement but none of lower Paleozoic age.
The CH-1 kimberlite has an area of close to 6 ha and consists of coherent (magmatic) kimberlite with pyrope garnet, chrome diopside, olivine phenocrysts up to 100 mm across, eclogite and peridotite xenoliths (Pell et al. 2008). It is exposed as cobbles in frost boils.

The CH-6 kimberlite underlies an area of ~1 ha and consists of carbonate-clast bearing kimberlite. Deeper in the pipe there are kimberlite facies that are either carbonate-poor or carbonate-free. Basement xenoliths are also rare. It is unclear whether this is a facies of the carbonate-xenolith kimberlite or whether it is a distinct and separate phase (Farrow et al., 2015). The CH-7 kimberlite, also about 1 ha, consists of two distinct lobes, the smaller of which consists of coherent kimberlite (CK) and the other of apparently coherent kimberlite (ACK) and volcanic kimberlite (VK) with clasts of carbonate and basement (Farrow et al., 2015). The CH-44 kimberlite has a surface area of 0.5 ha with ACK in the upper part and volcanic kimberlite (VK) or pyroclastic kimberlite (PK) at greater depths.

Qilalugaq

Exploration of the property by BHP Billiton from 2000 to 2005 included heavy mineral sampling, aerial and ground geophysics and drilling. This uncovered eight diamondiferous kimberlite pipes. The current owner is Stornoway Diamond Corp, and North Arrow Minerals in a joint venture, which has been continuing to bulk sample these pipes. Results point to a high proportion of coloured stones. The host rocks are mostly Archean granitoid gneiss and schist near the known pipes and, further afield, supracrustal rocks of the Archean Prince Albert Group unconformably overlain by Paleoproterozoic meta-sediments of the Penrhyn Group.

The kimberlite (emplacement age 546 Ma (late Ediacaran) is associated with an emplacement corridor 26 km long, 3 km wide and trending west-northwest. Within this corridor are 8 known pipes with surface areas ranging from 0.6 - 12.5 ha, and a dyke set. The Qilalugaq pipes are interpreted to be steep sided, pipe-like and irregular or elongate in plan view. The preserved material is diatreme-facies volcaniclastic kimberlite. The higher crater facies is missing and presumably eroded away.

The Q1-4 pipe is the primary focus of advanced exploration. Dominant near surface compositions include massive volcaniclastic kimberlite that grades laterally to tuffisitic kimberlite breccia and to depth grades to hypabyssal kimberlite. Multiple pipe-filling phases are found in some pipes, including Q1-4, in which the presence of five main filling phases has been documented. In addition, there are dykes and intrusions of hypabyssal kimberlite. Infill phases include:

1) A28a (41% of the pipe volume) consisting of volcaniclastic kimberlite grading downwards to hypabyssal kimberlite. Mineralogically this includes serpentinized olivine macrocrysts, matrix of serpentine, clinopyroxene, clay minerals, country rock xenoliths, minor phlogopite and perovskite. Mantle-derived indicators include ilmenite, garnet, chrome diopside and mantle peridotite.

2) The A48a phase, 14% of pipe volume consists of hypabyssal kimberlite.

3) A48b (20%) consists of volcaniclastic and hypabyssal kimberlite.

4) A61a (13%): hypabyssal kimberlite; and

5) A88a (12%): volcaniclastic kimberlite grading downwards to hypabyssal kimberlite.
GOLD PLACERS

Klondike

The Klondike placers are located in west-central Yukon in an area that has largely escaped glaciation (Figure 51). Initial discoveries in 1896 included gold placers on Quartz Creek, Gold Bottom Creek and most significantly, by George Carmack, Skookum Jim and Tagish Charlie, on Bonanza Creek which precipitated a massive staking rush.

Figure 51. Simplified geology of Yukon and western Northwest Territories showing placer gold mining districts.
Collectively the Klondike placers stand as one of the world’s most productive gold placers. The region has been more or less continuously productive for 119 years (1896–2015). Key placer streams include Hunker Creek, Bear Creek, Bonanza Creek, Dominion Creek, Eldorado Creek and surrounding drainage areas covering ca. 1200 km².

The Klondike area is unglaciated and therefore has not experienced glacial dispersion. This limits the lode gold source to several named ridges: Lone Star Ridge and Violet Ridge of which Lone Star has received the most significant attention. However, the gold content of these lodes does not account for the large volume of mined placer gold (Chapman et al., 2010).

The lode gold hosts are the Devonian to Mississippian and possibly older Nasina assemblage and the middle- to late-Permian (Guadalupian-Lopingian) Klondike Schist assemblage of the Yukon-Tanana terrane. These consist of greenschist-facies mafic and felsic metavolcanic rocks. Also present are thin slices of greenstone and ultramafic rocks of the Slide Mountain terrane. Together these are stacked into a series of thrust slices, subsequently eroded in mid- to Late Cretaceous time to produce locally derived sandstone, shale and conglomerate (Tantalus Formation). Associated volcanics of this age are assigned to the Carmacks Group. The later onset of motion on the Tintina Fault system is dated to early Eocene time. This fault zone remains active (Lowey, 2006; MacKenzie et al., 2008).

Placer gold occurs in four settings: 1) the lower part of the high-level White Channel gravels (5–3 Ma; up to 46 m thick) stratified with, and overlain by; 2) glaciofluvial Klondike gravel (Pliocene; up to 53 m thick), 3) intermediate terrace gravels of limited extent (1.4 Ma; up to 9 m thick), and 4) low-level gravels (late Pleistocene-Holocene; to 20 m thick) that occupy and underlie modern stream and river gravels at 10–200 m below the high-level gravels (Figures 52, 53; Lowey, 2006). Gold in the low-level gravels derives from erosion of both the White Channel gravels and primary bedrock sources.

Most of the placer gold is considered to be alluvial in origin. Some, however, may result from precipitation from water that carries gold in solution. Concerning the bedrock source, one
theory is that high-grade lode gold source(s) have been largely eroded away to form the present placer gold deposits leaving only lode gold relicts. An alternative is that one or more high-grade lode gold sources remain to be discovered.

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Deposits
- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Size and activity
- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Scale 1:9 500 000

Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 35°W
The Archaean

Greenland is mainly underlain by Archaean rocks in particular along the western coast and the eastern coast north to approximately 70°N, where the Archaean forms the basement of the Caledonian orogen in the south turning into Palaeoproterozoic basement further north (Enclosed map; Bridgwater et al., 1978; Garde and Steenfelt, 1999; Connelly and Mengel, 2000; Friend and Nutman, 2001; Thrane, 2002; Friend and Nutman, 2005; Dawes, 2006; Nutman et al., 2008a; Nutman et al., 2008b; Kolb et al., 2013; Kolb, 2014). The Archaean generally consists of two major cratons, the North Atlantic craton in the south and the Rae craton in the northwest with a possible continuation to the northeast and east of Greenland, separated by the broadly west-northwest-trending Palaeoproterozoic Nagssugtoqidian orogen (Enclosed map; van Gool et al., 2002; Connelly et al., 2006; Nutman et al., 2008a; Nutman et al., 2008b; Kolb, 2014). The Archaean is characterized by granulite-gneiss terranes or similar terranes at amphibolite facies metamorphic grades that host orogenic gold and orthomagmatic deposits (Kolb et al., 2015). The few lower-grade metamorphic areas locally host banded iron formation (BIF), such as those in the Melville Bay area and the Isua greenstone belt with the Isua iron ore deposit, located approx. 150 km northeast of the capital, Nuuk (Enclosed map; Appel, 1991; Dawes, 2006). The Isua iron ore deposit is isolated and not part of a province. Prospective BIF was, however, mapped over 19 km strike in the Melville Bay area (Figure 1), including coarse-grained magnetite and haematite BIF. The Havik deposit was found most prospective and an inferred resource of 67 Mt at 31.4 wt. % Fe magnetite iron ore was defined (Red Rock Resources, 2016).

The Palaeoproterozoic

Archaean rocks in central West and North-West Greenland are overlain by the Palaeoproterozoic Karrat Group, and both units are deformed by Palaeoproterozoic Rinkian-Nagssugtoqidian orogens (Enclosed map; Henderson and Pulvertaft, 1967; Garde and Pulvertaft, 1976; Henderson and Pulvertaft, 1987; Grocott and Pulvertaft, 1990; van Gool et al., 2002; Connelly et al., 2006). The Karrat Group is an approximately 8.5 km thick metasedimentary sequence metamorphosed at greenschist to lower amphibolite facies conditions with local higher grades in contact aureoles. It is subdivided into three formations, namely the Mârmorilik, Qeqertarssuaq and Nûkavsak formations (Garde and Pulvertaft, 1976). The Qeqertarssuaq and Nûkavsak formations are dominated by metamorphosed siliciclastic rocks, with the Nûkavsak Formation consisting of >1950 Ma metagreywacke deposits from juvenile Palaeoproterozoic and Archaean basement sources (Henderson and Pulvertaft, 1967, 1987; Grocott and Pulvertaft, 1990; Kalsbeek et al., 1998). The Mârmorilik Formation hosts the Black Angel Zn-Pb deposit (Pedersen, 1980, 1981), and is composed of calcite and dolomite marble, fine-grained, locally graphite-bearing schist, quartzite and possible metamorphosed evaporite (Garde and Pulvertaft, 1976). The Black Angel Zn-Pb deposit is part of a Zn-Pb district with numerous massive to disseminated sulphide targets in the vicinity (Thomassen, 1991). Southwest of the Black Angel deposit, the Lower Mârmorilik Formation hosts
mineralisation in east-southeast-trending fold hinges on the Ukkusissat peninsula, Appat island and Nuussuaq peninsula. The massive sulphide mineralisation on Ukkusissat peninsula is up to 8 m thick and 70 m long with 40 wt. % combined Zn+Pb and 72 ppm Ag (Thomassen, 1991). A massive sulphide mineralisation up to 1.5 m wide and consisting of pyrrhotite, sphalerite, galena and minor chalcopyrite is found 20 km north of the Black Angel deposit (Coppard et al., 1992). It is hosted in a 6-70 m thick marble horizon on top of Archaean orthogneiss. Massive sulphide ore occurs in lenses over 9 km strike with grades up to 41 wt. % Zn and 9.3 wt. % Pb over 1 m (Coppard et al., 1992). The various Zn-Pb sulphide ores in and around Black Angel as well as the mineralisation to the north, appear to be hosted by marble and controlled by fold structures, and thus form a Zn-Pb metallogenic province in central West Greenland. No other large metallogenic provinces are known from the Archaean and Palaeoproterozoic terranes.

The Mesoproterozoic to Cenozoic

South Greenland
The Mesoproterozoic Gardar intrusive suite in South Greenland formed in a period of continental rifting and alkaline magmatism lasting from ca. 1300 to 1140 Ma (Sørensen, 2006; Upton, 2013; Bartels et al., 2015). It consists of dyke swarms and 14 intrusive complexes with variable sizes from ~300 m up to 50 km in diameter (Upton and Emeleus, 1987). The alkaline and carbonatite magmas of the Gardar intrusive suite are enriched in Be, Li, F, Ga, P, Y, Zr, Nb, REE, Hf, Ta, U, Th, Fe, Ti, V and Zn. The degree of enrichment during magmatic differentiation, fractionation and hydrothermal alteration has determined their mineral potential (Steenfelt et al., 2016). This mainly rare metal province hosts some of the world’s largest deposits of this kind, namely Kringleerne (Ta-Nb-REE-Zr), Kvanejfeild (U-REE-Zn) and Motzfeldt (Nb-Ta) (Enclosed map). The central province is located around the settlements of Narsaq and Narsarsuaq and trends northeast. It is approximately 30 km wide and 120 km long. Further intrusions are found approximately 80 km both west and east. The Gardar intrusive suite is divided into (1) early Gardar Province with dominantly west-trending dykes and minor plutons (1300-1220 Ma) and (2) late Gardar Province with west-southwest-trending dykes and plutons (1180-1140 Ma) (Sørensen, 2006; Upton, 2013; Bartels et al., 2015). The magmas of the dykes and intrusive complexes are mildly to strongly alkaline and formed a long range of rock types, such as alkali-basalt, trachyte, comendite, phonolite and alkali-gabbro, syenite, alkali granite, nepheline syenite and carbonatite (Upton et al., 2003). The melt source for the various rocks of the Gardar intrusive suite is interpreted to be located in metasomatised sub-continenta lithospheric mantle (SCLM) above the garnet stability field (Goodenough et al., 2000; Marks et al., 2004; Upton, 2013; Bartels et al., 2015). Minor degrees of crustal contamination, magmatic differentiation by crystal fractionation and volatile transfer concentrated rare metals in the most evolved magmas and magmatic-hydrothermal fluids, but also formed magnetite or eudialyte cumulate. Most of the ores are hosted in the plutons themselves, with the exception of the Motzfeldt deposit where the magmatic-hydrothermal fluids altered the wall rocks and formed pyrochlore mineralisation.

Southern West Greenland
Southern West Greenland has remained tectonically relatively stable since the Palaeoproterozoic and has only sporadic records of younger rock until Cretaceous-Palaeogene rifting formed sedimentary rocks and flood basalts around 70°N (Enclosed map). The Archaean and Palaeoproterozoic gneisses host, locally, carbonatite complexes and dyke swarms described as ultramafic lamprophyre and kimberlite (Larsen, 1991; Nielsen et al., 2009). This carbonatite-ultramafic lamprophyre province is located at the southern front of the Palaeoproterozoic Narsaqgtoqidian Orogen (Enclosed map; Secher and Larsen, 1980; Scott, 1981; Nielsen et al., 2009). The ~90 km² and ca. 565 Ma Sarfartoq carbonatite complex forms a conical body with peripheral radial dykes and is situated approx. 50 km southwest of the Kangerlussuaq international airport (Enclosed map). This complex hosts a world class REE mineralisation, the Sarfartoq deposit (Druecker and Simpson, 2012). Dyke swarms of various ages are widely distributed in the Sisimiut-Sarfartoq-Maniitsoq region and locally host diamond, e.g. at Garnet Lake (Hutchison and Frei, 2009). They generally form ~1 m thick sheets
or 1-2 m wide near-vertical dykes in approx. 50 km wide halos around the carbonatite. The geometry of the dykes and sheets is controlled by the foliation of the host gneiss: 1284-1209 Ma lamproite dykes in the Sisimiut area (Rb–Sr phlogopite ages) trend east to east-southeast, whereas the 604-555 Ma (Rb–Sr phlogopite and U–Pb perovskite/pyrochlore ages) dykes trend east-northeast in the Maniitsoq area and northward in the Sarfartoq area (Larsen and Rex, 1992; Secher et al., 2009).

Northern Greenland

In northern Greenland, the large Franklinian basin developed during the Neoproterozoic and Palaeozoic at the margin of Laurentia with a ~2000 km E-W extent into Canada (Enclosed map; Higgins et al., 1991). The Neoproterozoic-Palaeozoic rocks unconformably overlie Proterozoic sedimentary sequences (Higgins et al., 1991; Dawes, 2009). Sedimentation started in the Neoproterozoic and lasted until the Early Devonian, when the basin was inverted during the Ellesmerian orogeny (Higgins et al., 2000). The strata in the Franklinian basin are characterized by a distinct facies transition from ~3 km thick shallow-water platform carbonates in the south to ~8 km thick deep-water, mainly fine-grained siliciclastic trough sediments separated by the Navarana Fjord lineament in the north (Soper and Higgins, 1990; Higgins et al., 1991). In later stages of basin evolution, thermal subsidence controlled sedimentation in a starved basin until the Silurian (Hopper and Higgins, 1987). Carbonate, carbonate conglomerate and black shale were deposited on the shelf slope, and cherty shale formed continuously in the trough until the end of the Ordovician (Higgins et al., 1991). Shelf carbonate formed until the Llandovery, followed by turbidite deposition from Wenlock to Early Devonian times (Soper and Higgins, 1990; Higgins et al., 1991). During this period, the platform-trough transition migrated southwards, which is explained by the onset of the Ellesmerian orogeny in the north (Soper and Higgins, 1990). The first indication of mineralisation in the region was reported in 1969 during the Skidoo British Joint Services Expedition followed by mapping of the Geological Survey of Greenland in the years 1978-1980. Several Zn-Pb sulphide occurrences along the Navarana Fjord lineament may represent SEDEX occurrences similar to Citronen, defining a metallogenic Zn-Pb belt in northern Greenland (von Guttenberg and van der Stijl, 1993). These include the showings in Nares Land (Kap Wohlgemuth) and Nyboe Land (Hand Bugt and Repulse Havn) with up to 11 wt. % Zn and 2 wt. % Pb in grab samples (von Guttenberg and van der Stijl, 1993). Further Zn-Pb occurrences to the south are hosted in the shelf carbonates and probably represent MVT-style mineralisation that formed during Ellesmerian basin inversion (Rosa et al., 2013; Rosa et al., 2014).

East Greenland

East Greenland is, in its northern part, characterized by variably deformed and metamorphosed rocks that mainly formed between latest Palaeoproterozoic and Silurian times. Sedimentary rocks with local volcanic intercalations were deposited in continental to shallow marine basins along the eastern margin of Laurentia, and overlying the Archaean and Palaeoproterozoic basement (Enclosed map; Henriksen and Higgins, 2008). The Caledonian orogen resulted in the formation of regional-scale west-vergent nappe stacks and local granite intrusions (Leslie and Higgins, 2008). The Devonian marks the end of Caledonian deformation and the development of sedimentary basins in central East Greenland, before continental break-up and initiation of sea-floor spreading in the North Atlantic started in the Palaeogene. Post-Caledonian sedimentation up to the Cretaceous occurred in isolated north-trending basins that form graben structures. The sedimentary rocks locally host Cu-mineralisation and Caledonian granites formed a number of skarn or greisen occurrences (Harpoon et al., 1986).

The continental break-up and onset of Palaeogene sea-floor spreading closely related to a mantle plume was associated with major magmatism resulting in massive lava flows, intrusions, sills and dykes (White and McKenzie, 1989; Saunders et al., 1997). This large igneous province (LIP), the North Atlantic Igneous Province (NAIP), has been active since 63 Ma and extends for more than 3000 km from Baffin Island across
Greenland, Iceland and the Faeroe Islands to Scotland and Ireland. The Palaeogene in East Greenland stretches from 66°N to 75°N, a distance of more than 1300 km (Enclosed map). Thick volcanic sequences also exist off-shore on the eastern Greenland shelf (Brooks and Nielsen, 1982; Larsen et al., 1989; Pedersen et al., 1997). Evidence of the magmatism is also preserved as dyke swarms, sill complexes, gabbroic intrusions and alkaline and granite intrusions. The 63-13 Ma NAIP in eastern Greenland was formed by episodic magmatism (Tegner et al., 1998; Tegner et al., 2008; Larsen et al., 2014). The three main magmatic stages are: (1) ca. 62-57 Ma picritic and tholeiitic volcanism (Lower Basalts) over a wide area, possibly triggered by plume activity; (2) ca. 56-55 Ma tholeiitic volcanism forming flood basalts (Plateau Basalts) and intrusions related to continental rifting; and (3) ca. 53-47 Ma gabbro and syenite intrusions and transitional alkaline volcanism in a late post-breakup phase. Sporadic younger magmatism is locally recorded at Kialineq at ca. 35 Ma, Malmbjerg at ca. 26 Ma and finally by the small volumes of the ca. 13 Ma-old Vindtoppen Formation basalt. More than 60 intrusions are found along the eastern Greenland margin, with 28 in the Kangerlussuaq Fjord area alone. The mafic to ultramafic intrusions host orthomagmatic PGE-Au mineralisation, the largest deposit known being in the Skaergaard intrusion (Holwell et al., 2012a; Nielsen et al., 2015). The felsic intrusions include both undersaturated and oversaturated syenite and monzogranite, and host Mo-porphyry ores as well as peripheral Au-rich veins and Zn-Pb vein mineralisation, e.g. at Malmbjerg and Flammefjeld (Schassberger and Galey, 1975; Geyti and Thomassen, 1984). Volumetrically minor alkaline intrusions form the Gardiner Complex, at the head of the Kangerlussuaq Fjord, and carbonatites of the Sulugssut Complex (Brooks, 2011). The Kangerlussuaq basin and the Kangerlussuaq Fjord form the best endowed area, with orthomagmatic PGE-Au, Mo-porphyry and Au-Ag vein deposits (Skaergaard, Flammefjeld, Kap Edward Holm, macrodykes) and is interpreted as the locus of the mantle plume (Enclosed map; Brooks, 1973; Schassberger and Galey, 1975; Brooks and Nielsen, 1982; Geyti and Thomassen, 1984; White and McKenzie, 1989; Arnason and Bird, 2000; Thomassen and Krebs, 2001; Holwell et al., 2012a; Nielsen et al., 2015). The Kangerlussuaq Fjord is interpreted to represent the failed arm of a triple junction with break-up and rift-related magmatism concentrated in two arms forming the present Greenlandic coast line (Brooks, 1973).

HISTORY OF MINING
Jochen Kolb

Greenland has a long tradition of mining and mineral exploration starting with the colonisation in the 18th century with a first culmination marked by the discovery and subsequent mining of a cryolite (Na₃[AlF₆]) deposit near Ivittuut in South Greenland. Cryolite was used first as an aluminium ore and then as flux in the electrolytic processing of aluminium from bauxite ore. The Ivigtut cryolite mine was mined for over 130 years from 1854 until 1987 and represents one of the few economic mining successes in Greenland so far (Pauly and Bailey, 1999). The mine was the only one of its kind in the world until natural cryolite was replaced by synthetic cryolite in the 1960s. The company that operated the Ivigtut cryolite mine, Kryolitselskabet Øresund A/S, started the first systematic exploration program in the 1950s with helicopter prospecting flights mainly along the west coast, selecting targets for mapping and further exploration. At the same time, Nordisk Mineselskab A/S started mineral exploration programs in eastern Greenland based on finds from earlier expeditions, i.e. the small Blyklippen Zn-Pb deposit that was mined between 1956 and 1962 (Harpøth et al., 1986). Other early, small mining operations include the Josva copper mine (1904-1915) and the Amitsoq graphite mine (1915-1924) in
South Greenland, and the Qullissat coal mine (1924-1972) on Disko Island in central West Greenland. The discovery of the world-class Zn-Pb Black Angel deposit is the second important event in Greenland’s mining history. The deposit was successfully mined between 1973 and 1990 in a spectacular operation, where ore and the miners were transported between the adit at an elevation of 600 m above sea level and the opposite side of a fjord by a cable car. Several exploration companies became active in the wake of the successful mining operations, identifying several mineral exploration targets mainly by prospecting. Mineral exploration, geophysical and geological investigations have been supported through the partly state-owned exploration company Nunaoil A/S with the split-off NunaMinerals A/S and the geological surveys (Geological Survey of Greenland and later Geological Survey of Denmark and Greenland). This resulted in the discovery of gold and olivine deposits, which led to establishment of the Nalunaq gold mine (2003-2013) in South Greenland and the Seqi olivine mine (2005-2010) north of Nuuk, respectively. With the termination of gold production in Nalunaq in 2013 and the final closure of the mine in 2014, mining activity came to halt in Greenland for the first time in 160 years. Status January 2016, approximately 55 exclusive exploration and 6 exploitation licenses are active, covering approximately 10 % of Greenland’s ice-free area (Government of Greenland, 2016). Exploration activity focuses on the Proterozoic of South Greenland (REE, Fe-Ti, Au, U), the Archaean between Fiskenæsset and Sisimiut in western Greenland (Cr, Ni, PGE, Ti-V, REE, gemstones), the Archaean and Palaeoproterozoic in northwestern Greenland (Zn, Fe, Au, Cu) and the Caledonian Orogen of eastern Greenland (Cu, W). Exploitation licenses have been granted for the Nalunaq gold deposit, the Black Angel zinc-lead deposit, the Malmbjerg molybdenum-porphry deposit, the Isua iron ore deposit, the Aappaluttoq ruby deposit and the White Mountain calcium feldspar deposit. Mining of corundum gemstone from the Aappaluttoq ruby deposit commenced in December 2015. Reviews of the economic geology of Greenland are given in Ball (1922), Nielsen (1973, 1976), Henriksen et al. (2009) and Kolb et al. (2015); Kolb et al. (in review).

ISUA IRON ORE DEPOSIT

Jochen Kolb

The Isua iron ore deposit was discovered in 1965 by Kryolitselskabet Øresund A/S during an aeromagnetic survey 150 km northwest of Nuuk (Nielsen, 1973). It was studied in detail with drilling and the definition of a resource of magnetite ore until 1974, but was considered only marginally economic. Exploration was resumed in 1995, when RTZ Mining and Exploration Ltd. started investigating haematite ore targets underneath the Inland Ice (Figure 1, Enclosed map; Coppard and Harris, 1996). Recent resource estimates on re-examined magnetite ore, calculated by London Mining in 2014, gave figures of 1107 Mt grading 32.6 % Fe. General Nice Development Limited took over the license for the Isua iron ore deposit in 2015.

The deposit is hosted in an oxide-facies banded iron-formation (BIF) in the eastern part of the Isua greenstone belt in the northern ca. 3700 Ma terrane of the Isua greenstone belt (Nutman and Friend, 2009). The BIF crops out as a sheet, 1.5 km long, 200 m thick and dipping east-southeast at 60-70°. it is located along the hinge of a large-scale antiform close to the Inland Ice and increases in thickness to 450 m with depth (Figures 1 and 2). The lower section is formed by iron-rich garnet-chlorite biotite schist and the upper section by a laminated quartz-carbonate-magnetite-iron silicate rock with 20-30 wt. % Fe (Keto, 1970). The BIF is laterally zoned with magnetite-rich rocks in the centre grading to more silicate-rich rocks in
the periphery. The quartz- and magnetite-rich bands are 0.5-100 mm thick (Nielsen, 1973; Appel, 1980; Frei et al., 1999). The quartz-rich bands consist additionally of minor chlorite, calcite, tremolite and grunerite, whereas the magnetite-rich bands contain minor pyrite, tremolite, quartz, chalcopyrite and pyrrhotite (Keto, 1970). The banding is folded and an axial planar foliation defined by amphibole and magnetite is developed at high angles to the banding (Frei et al., 1999). Magnetite is replaced by haematite (martitisation) in areas of strong folding. Locally, 10 cm wide L-type mylonite includes cigar-shaped composite aggregates of magnetite, actinolite and quartz, and 10-15 cm wide magnetite-quartz brecciae with a magnetite matrix (Appel, 1980). Quartz-calcite-chlorite veins crosscut the banding (Keto, 1970).

Only loose boulders of haematite ore are found in ice-free areas, which have haematite- and quartz (-magnetite) bands or deformed haematite bands in a yellow jasper matrix (Appel, 1980). The haematite BIF is additionally composed of sericite, quartz and minor relic magnetite and contains ~60 wt. % Fe and ~14 wt. % Si (Coppard and Harris, 1996). In drill cores, the haematite BIF has drusy structures that are filled by quartz.
and is crosscut by quartz-jasper veins. The drilled occurrence of haematite BIF has a lower fault contact at 678 m and is underlain by magnetite BIF (Coppard and Harris, 1996).

The mineralisation formed as an Algoma-type BIF at 3691 ± 22 Ma due to hydrothermal fluids percolating through mid-ocean ridge-type rocks (Frei et al., 1999). Quartz-magnetite-grunerite-stilpnomelane or quartz-magnetite-actinolite-grunerite-ferrosalite assemblages indicate amphibolite facies metamorphism (Frei et al., 1999). Discordant veins with ~5 mm hydrothermal alteration zones formed during the amphibolite facies metamorphism at 3630 ± 70 Ma (Nielsen, 1973; Frei et al., 1999). Magnetite is locally replaced by fine-grained haematite, indicating alteration of the magnetite BIF (Nielsen, 1973; Appel, 1980). The haematite ore probably represents an upgraded ore formed by oxidation of the magnetite BIF during hydrothermal or supergene alteration.

Figure 2. Photograph of the Isua BIF outcrop close to the Inland Ice during a core drilling campaign (C. Østergaard).

BLACK ANGEL ZINC-LEAD DEPOSIT
Diogo Rosa

The Black Angel deposit, at Maarmorilik, comprises ten distinct ore bodies: Angel, Cover, Tributary, I and I South, Banana, Deep Ice, V16 and Nunngarut 1 and 2 (Figure 3; Enclosed map). These totalled 13.6 Mt, with grades of 12.3 % Zn, 4.0 % Pb and 29 ppm Ag (Thomassen, 1991), of which 11.2 Mt were mined between 1973 and 1990. Upon closure of the mine, approximately 2.4 Mt of ore remained in pillars, which have been planned to be exploited. The massive sphalerite-galena-pyrite ore is hosted by calcitic and dolomitic marble (Figure 3), with intercalations of anhydrite-bearing marble and pelitic schist, of the Mârmorilik Formation of the Palaeoproterozoic Karrat Group. The Karrat Group corresponds to a meta-sedimentary sequence which unconformably covers the Archaean basement in northern West Greenland (Henderson and Pulvertaft, 1987). Towards the south, the Karrat Group consists mostly of the carbonate-dominated Mârmorilik Formation. Towards the north, the Karrat Group includes the lower, mainly silicilastic (quartzitic), Qeqartarsuaq Formation, and the upper, turbiditic Nûkasvsak Formation, which contains occasional pyrrhotite-chert-graphite horizons. Amphibolites present between the two latter formations have been interpreted to represent a volcanic package, referred to as the Kangigdleq basic volcanics by Allen and Harris (1980).

The genesis of the ores, whether sedimentary-exhalative or a later stage Mississippi Valley-type, remains uncertain (Thomassen, 1991) because of poor age constraints and the effects of deformation, as well as the 1881±20 Ma amphibolites-facies metamorphism (Taylor and Kalsbeek, 1990). These processes during the evolution of the Rinkian mobile belt, caused the original sulphide sheets, which were originally 0.5 to 8 m thick, to be deformed into 0.5-35 m thick, flat-lying sheets along a 3 km long and up to 600 m wide zone (Pedersen, 1980, 1981).
Figure 3. (a) Schematic map showing the Black Angel Zn-Pb deposit and its various ore bodies. The inset shows Zn-Pb mineralisation in the district. (b) Sketch of the ore body and marble host rock showing the intense ductile deformation defining the deposit geometry. (c) Schematic cross section through the Black Angel mine (modified after: Thomassen, 1991).
ISORITOQ IRON-TITANIUM-VANADIUM DEPOSIT
Kristine Thrane

The Isortoq deposit occurs in a troctolite dyke which intrudes the Julianehåb igneous complex in the Qaortoq District of South Greenland, approximately 70 km west of Narsaq. The dyke extends over 16.3 km, and is one of the many Mid-Proterozoic mafic dykes in the area (Bridgwater and Coe, 1970). The deposit is referred to as the Isortoq Project and is licensed to West Melville Metals Inc. In 2012, the company carried out a ground geophysical survey and drilled 11 holes, producing a total of 2,684 m of core. The drilling program was carried out along strike for approximately 1 km of the mineralised body, and showed that its width ranges from 100 to 200 m and its vertical thickness from 100 to 300 m. In 2013, West Melville Metals Inc. announced an initial National Instrument (NI) 43-101 compliant resource estimate of 70.3 Mt grading 38.1 wt. % FeO (29.6 wt. % Fe), 10.9 wt. % TiO\(_2\) and 0.144 wt. % V\(_2\)O\(_5\) applying a 15 wt. % Fe cut-off.

Bench-scale beneficiation tests carried out in 2014 produced average concentrate grades of 50.2 wt. % Fe (71.8 wt. % Fe\(_2\)O\(_3\)); 20.9 wt. % TiO\(_2\) and 0.34 wt. % V\(_2\)O\(_5\) with acceptably low levels of penalty elements, such as S, P, Si, Al (West Melville Metals Inc., 2016).

MOTZFELDT NIOBIUM-TANTALUM-ZIRCONIUM-RARE EARTH ELEMENT DEPOSIT
Kristine Thrane

An extensive Nb-Ta-U-Th-Zr-REE mineralisation was discovered in the centre of the Motzfeldt intrusion during the reconnaissance surveys of the Syduran project (Bridgwater and Coe, 1970). The Motzfeldt intrusion is part of the Mesoproterozoic Igaliko alkaline intrusive complex in the Gardar Province (Figure 4, Enclosed map). It consists of multiple units of syenite and nepheline syenite that intruded the Proterozoic Julianehåb intrusive complex and the unconformably overlying supracrustal rocks, the Eriksfjord Formation. The Motzfeldt intrusion covers an area of approximately 150 km\(^2\). McCreath et al. (2012) dated the intrusion at 1273 ± 6 Ma, by concordant U-Pb zircon ages. The central part of the Motzfeldt intrusion, called Motzfeldt Centre, is divided into the Geologfjeld, Motzfeldt So and Flinks Dal Formation, where each part is divided into several subdivisions (Figure 4). At least two sets of faults cut the intrusion; the older generation of faults strikes SW-NE and those of the younger generation strike E-W. The coarser syenitic rocks are intruded by sills or sheets of peralkaline microsyenite in the north and north-eastern part of the Motzfeldt Centre.
The Nb-Ta-U-Th-Zr-REE mineralisation is hosted by the altered syenites and microsyenites of the Motzfeldt So Formation. The metals are mainly hosted in pyrochlore (Nb, Ta, U, REE), thorite (Th), zircon (Zr), bastnaesite (REE, Th) and monazite (REE) (Thomassen, 1988, 1989; Lewis et al., 2012). The pyrochlore is concentrated into several 100 m. The width of most of the zones lies in the range of several to >100 metres. The pyrochlore contains 3-9 wt. % UO_2 and up to 0.25 wt. % ThO_2 (Tukiainen, 1986). Significant Ta-Nb-enriched zones are also related to minor pegmatite and diorite dykes and high-grade REE intersections are related to pegmatites.

Regency Mines Plc. currently holds the license for the area. The resource estimate by the former licence holder, RAM resources Ltd, is 340 Mt ore grading 0.26 wt. % TREO, 0.19 wt. % Nb_2O_5, 0.012 wt. % Ta_2O_5 and 0.46 wt. % ZrO_2.

Figure 4. Geological map of Motzfeldt intrusion (modified after: Tukiainen et al., 1984). The great numbers of dykes, predominantly alkali trachytes of the late-Gardar dyke swarm are omitted for the sake of clarity.
The Ivigtût intrusion is one of the most evolved complexes in the Gardar Province, but it is the only one that intruded Archaean basement. A Rb-Sr isochron yields an age of $1248 \pm 25$ Ma for the granite (Thomassen, 1989). The granite stock is 300 m in diameter and expands with increasing depth. The unique cryolite ($\text{Na}_3\text{AlF}_6$) body in the intrusion was mined from 1854 to 1987 (Figure 5). In addition, an occurrence of uranium was discovered by Kryolitselskabet Øresund A/S when carrying out radiometric investigations around the cryolite mine. The main U occurrence is associated with columbite in the marginal areas of the cryolite body. The main cryolite body is now exhausted, and the mine abandoned; the majority of the remaining tailings stored adjacent to the mine site have also been processed. A halo of radioactivity around the Ivigtût intrusion has been recognized and investigated; it contains 50-100 ppm U in the highly altered granites. The U/Th ratio is approximately 1:4 (Pauly, 1960).

Rimbal Pty. Ltd. currently holds the license for the area, exploring for ultra-pure quartz beneath the former cryolite body.

Figure 5. Abandoned mine site at Ivigtût.
The 150 km² Mesoproterozoic Ilímaussaq alkaline complex hosts the REE-U-Th-Zn-F deposit referred to as Kvanefjeld (Figure 6, Enclosed map). The complex is unique and world famous for its wealth of rare minerals, several of which are only found at Kvanefjeld. It has been studied intensely since the beginning of the 19th century for both scientific and economic reasons (see Sørensen, 2001) for an overview. Ilímaussaq is c. 1160 Ma old (Lewis et al., 2012) and one of the youngest

Figure 6. Geological map of the Ilímaussaq intrusion and a schematic cross-section (modified after Upton, 2013). The REE deposits hosted by lujavrite indicated.
intrusions of the Gardar igneous province. It intruded the Palaeoproterozoic Julianehåb igneous complex and the unconformably overlying Mesoproterozoic Eriksfjord formation comprising sandstone and basalt. The complex is composed of a series of nepheline syenites and is the type locality for agpaitic rocks. Kvanefjeld represents the intermediate series of the Ilímaussaq intrusion, sandwiched between the roof and floor series, and is composed of hyper-agpaitic lujavrite and nauja rite. The lujavrite represents the residual liquid after the crystallization of the highly differentiated alkali- and volatile-rich magma with high concentrations of incompatible elements (Bohse et al., 1974). Kvanefjeld has an average $U_3O_8$ concentration of 273 ppm and approximately 3 times the amount of Th. In addition, it has an average TREO concentration of 1.06 wt. % (402 ppm HREO) and 0.22 wt. % Zn. The REE, U and Th are predominantly hosted by the complex phosphate-silicate mineral steenstrupine, containing 0.2-1.5 wt. % U and 0.2-7.4 wt. % Th.

Kvanefjeld was explored for uranium between 1956 and 1984. Geological mapping and radiometric surveys were carried out, 12,455 metres of core were drilled and a 1 km long adit was constructed. Since 2007, Greenland Minerals and Energy Ltd. has conducted REE-exploration activities in the Kvanefjeld area, including drilling of an additional 57,710 m of core. Greenland Minerals and Energy Ltd. reports (2016) that the overall resource inventory for Kvanefjeld (150 ppm $U_3O_8$ cut-off) is 673 Mt of ore containing 167,000 t $U_3O_8$ and 7.34 Mt TREO, including 0.247 Mt HREO, and 1.36 Mt Zn. An application for an exploitation licence is expected to be submitted early in 2016. Planned annual production is 3 Mt ore, equal to 23,000 t TREO, 500 t $U_3O_8$ and 6,000 t Zn. Additional resources exist in Zone Sørensen and Zone 3.

**Kringlerne Tantalum-Niobium-Rare Earth Element-Zirconium Deposit**

Kristine Thrane

The Kringlerne multi-element deposit is hosted in the basal cumulates, the kakortokites, of the Ilímaussaq alkaline complex (Figure 6, Enclosed map). The kakortokites crop out in the southern part of the Ilímaussaq complex, and consist of 29 exposed cyclic units with a total thickness of about 200 m (Bohse et al., 1971). Each individual unit is made up by 3 density-stratified layers: a dark arfvedsonite dominated kakortokite layer, a reddish eudialyte dominated kakortokite layer, and a grey/white nepheline-feldspar dominated kakortokite layer (Figure 7). Eudialyte, the main economic mineral, is enriched in REE-Zr-Nb-Ta. Rimbal Pty Ltd., the current license-holder reports an inferred resource to be at least 4,300 Mt, grading 0.65 wt. % TREO, 0.2 wt. % Nb$_2$O$_5$ and 1.8 wt. % Zr$_2$O$_5$ equalling 28 Mt TREO (Rimbal Pty Ltd., 2016). An application for an exploitation licence has been submitted and is currently under review. The proposed mining project involves an open-pit mine near the fjord, hauling ore to a nearby beneficiation plant, where three products will be produced by magnetic mineral separation: (i) eudialyte concentrate (REE, Nb, Zr); (ii) feldspar concentrate; and (iii) arfvedsonite concentrate, all to be shipped for further processing or use outside Greenland. Planned annual production is 500,000 t ore, equal to 3,250 t TREO (equivalent to 400 t Nd$_2$O$_3$ and ca. 90 t Dy$_2$O$_3$) and 9,000 t Zr$_2$O$_5$.
The ca. 565 Ma old Sarfartoq carbonatite complex contains 14 Mt ore grading 1.5 wt. % total rare earth oxides (TREO) with a cut-off at 0.8 wt. % TREO (Druecker and Simpson, 2012). It has been explored since 2009 by Hudson Resources, and 25,400 m of core have been drilled and 273 km of geophysical surveys have been flown. Mineral exploration in the 1980’s and 1990’s by Hecla Mining and New Millennium Resources NL concentrated on a small high-grade Nb resource near the core of the complex. The resources to a depth of 90 m are estimated at 64,301 t at 3.89 wt. % Nb corresponding to 5.56 wt. % Nb$_2$O$_5$ at a cut-off grade of 1.0 wt. % Nb (Woodbury, 2003; Stendal et al., 2005).

The carbonatite complex consists of concentric intrusions of rauhaugite and beforsite in the up to 4 km wide and 5 km long intrusion centre forming a conical body with an axis, which plunges steeply to the northwest (Figures 8 and 9, Enclosed map). This centre is surrounded by a fenite zone several hundred metres wide and a marginal zone, which is several kilometres wide. The rauhaugite in the intrusion centre consists of dolomite, apatite, biotite, magnetite, richterite-arfvedsonite solid solution and minor zircon, ilmenite, pyrochlore and pyrite (Secher and Larsen, 1980). The concentric intrusions are 0.5-20 m wide, strongly foliated and layered, and host abundant wall-rock xenoliths. The layering is defined by local magnetite and beforsite bands in larger intrusions. The dyke- and sheet-like intrusions are separated by narrow bands of fenitized wall rock, which also forms the xenoliths (Figure 9c). The structures vary from foliated
Figure 8. Geological map of the Sarfartoq carbonatite complex, showing the conical intrusion centre, the narrow fenite rim and the kilometre-wide marginal zone (modified after: Secher and Larsen, 1980).

Figure 9. (a) Panorama view of the Sarfartoq carbonatite in reddish colours. The altered wall rock orthogneiss is seen with grey colours at the right and left margin (J. Lautrup). (b) View of drill rig and the carbonatite in the cliff wall. In the distance the wall rock shows shallow-dipping bands of orthogneiss and amphibolite. The plateau is at an elevation of approximately 800 m above sea level. (c) Rauhaugite intrusion with bands of fenite from the intrusion centre (L.M. Larsen).
to folded and brecciated, indicating a dynamic setting for emplacement of the carbonatite. The fenites are aegirine-bearing orthogneiss and generally contain alkali feldspar, aegirine, biotite, amphibole and minor calcite and pyrochlore (Secher and Larsen, 1980).

The marginal zone is characterized by haematite alteration and 50-200 m wide cataclastic fault zones that have a broadly radial geometry (Secher and Larsen, 1980). Monomineralic pyrochlore veins, 1-5 m wide also form a radial geometry surrounding the central conical intrusion. The marginal zone also hosts radial beforsite dykes, 5-100 cm wide and concentric breccia zones with carbonatite matrix. The beforsite consists of the same assemblage as the rauhaugeite with additional K-feldspar, chlorite, quartz, baryte, pyrite, haematite and rutile (Figure 8; Secher and Larsen, 1980). Late hydrothermal carbonate and calcite-fluorite veins cross-cut the carbonatite complex.

The REE mineralisation is hosted by the beforsite (ferrocarbonatite) dykes in the periphery of the complex and was identified by a radiometric Th-anomaly during exploration (Secher and Larsen, 1980; Druecker and Simpson, 2012). In the beforsite, the REE mineralisation is mainly located in bastnaesite, monazite, synchysite (CaCe(CO$_3$)$_2$F) and zhonghuacerate (Ba$_x$Ce$_{0.8}$La$_{0.2}$Nd$_{0.1}$(CO$_3$)$_3$F). The REE-bearing carbonates are enriched in Ce, La, Nd, Pr and Eu, yielding ore that is dominated by the light REE (Secher and Larsen, 1980; Druecker and Simpson, 2012).

The Sarfartoq carbonatite complex formed in two major stages: (1) intrusion of a central conical body; and (2) intrusion of concentric and radial beforsite dykes in the periphery (Secher and Larsen, 1980). The mineralisation formed in the late stages of carbonatite magmatism by subsolidus REE-enrichment in hydrothermal fluids associated with the peripheral dykes (Secher and Larsen, 1980).

Citronen Zinc-Lead Deposit

Diogo Rosa

Citronen is a large sedimentary-exhalative deposit located in Peary Land, in northern Greenland (Figure 10; Enclosed map), with reported total resources (measured + indicated + inferred), at 2.0 wt.% Zn cut off, of 132 Mt @ 4.0 wt.% Zn and 0.4 wt.% Pb (Ironbark Zinc Limited, 2012). This yet unmined deposit is composed of five major mounds, located within three main mineralised horizons (Figure 10). Each stratabound sulphide mound has dimensions reaching a length of 1500 m, a width of 600 m, and a thickness of 25 m, with individual sulphide beds being 0.3 to 2.0 m thick and interlayered with thin (≤10 cm) beds of mudstone (Kragh et al., 1997). The mounds consist of massive and bedded pyrite, with variable amounts of sphalerite and galena, and are hosted by mudstone, shale and debris flows of the Ordovician Amundsen Land Group (van der Stijl and Mosher, 1998). This group was deposited in a sediment-starved trough or elongate basin, separated from the shallow-water carbonate platform to the south by a prominent paleo-escarpment, the Navarana Fjord escarpment (Higgins et al., 1991; van der Stijl and Mosher, 1998). It has been suggested that the local basin bottom waters shifted from oxic to anoxic and locally sulphidic. This was caused by the emplacement of debris-flow conglomerates that sealed off the basin from oxic seawater, and due to the venting of reduced hydrothermal fluids, which allowed for the preservation of sulphides on the sea floor (Slack et al., 2015). Subsequent deformation is expressed by open to recumbent folds and local thrust faults, related to the Late Devonian to early Carboniferous Ellesmerian orogeny (van der Stijl and Mosher, 1998).
The layered mafic Skaergaard intrusion, located at the mouth of the Kangerlussuaq fjord in East Greenland (Figure 11, Enclosed map), hosts a major stratabound PGE-Au mineralisation with a total resource of 202 Mt at 0.88 g/t Au, 1.33 g/t Pd and 0.11 g/t Pt (Platina Resources Ltd., 2013). In addition to the precious metals, the deposit is rich in oxides and could be of interest for multi-element extraction of Ti, V and Fe. The deposit was discovered by Platinova Resources Ltd. in 1987 and has subsequently been explored by Skaergaard Minerals Corp. and Platina Resources Ltd. The three companies have drilled 68 diamond boreholes in total, corresponding to more than 35 km.

The intrusion is world famous for its spectacular layering and has served as a natural laboratory for studying magma chamber processes and differentiation of basaltic liquids. Extensive descriptions and reviews on the intrusion are given by Wager and Brown (1968); McBirney (1996); Irvine et al. (1998). The Skaergaard intrusion was emplaced at 56.0 Ma (Wotzlaw et al., 2012) into a fault-bounded space at the unconformity between the Precambrian basement and the overlying Eocene flood basalts. The intrusion is part of the major Eocene magmatic episode associated with the early continental break-up during the opening of the North Atlantic. The intrusion, which measures...
approximately 7 x 11 km (Figure 11), crystallised under closed-system conditions from ferrobasalt magma and developed through a continuous crystallisation sequence from olivine gabbro to ferrodiorite. It is divided into three zones originally defined by Wager and Deer, (1939): the Layered Series (LS), the Upper Border Series (UBS), and the Marginal Border Series (MBS), which crystallised on the floor, roof and walls of the chamber, respectively (Figure 12a). The interface at which the downward crystallising UBS met the upward crystallising LS is known as the Sandwich Horizon (SH). Each of the three series are further subdivided into Lower-, Middle-, and Upper Zone and subzones based on phase layering (Figure 12a, e.g. Wager and Brown, 1968; Irvine et al., 1998). Extreme fractional crystallisation is reflected in the cryptic variation resulting in pure iron end-members for olivine and Ca-rich pyroxene, and the progressive change in plagioclase, without reversals, from An70 at the margins to An25 at the centre of the intrusion (McBirney, 1996).

The upper part of the Middle Zone of the Layered Series hosts a large PGE and gold mineralisation known as the Platinova Reef (Figures 11 and 12; Bird et al., 1991) with maximum grades of 220 ppb Pt, 5.1 ppm Pd and 20 ppm Au (Bird et al., 1991; Andersen et al., 1998). The reef is hosted in the Triple Group which is a 100 m thick macro-rhythmic unit of three characteristic leucocratic layers (L1-L3) that are easily identified in drill cores as well as in the field (Figure 13). A buff-coloured, thinner and less distinct layer 20 m below L1 has been accepted as a member of the Triple Group and it is referred to as L0 and defines the base of the reef (Nielsen et al., 2015). The mineralisation, which has a total thickness of c. 50 m, is perfectly concordant with the L-layers and is defined by five main Pd-rich layers and one Au-rich layer. It consists of a stack of compositionally zoned, saucer-shaped layers of decreasing diameter, separated from each other by a layer of non-mineralised ferro-diorite of constant width (Figure 12b). The sulphides are entirely dominated by bornite, digenite and chalcocite while the precious metals are alloys of palladium, gold and copper, with skaergaardite (PdCu), nielsenite (PdCu3), zviagintsevite (Pd3Pb) and tetra-auricupride (AuCu) being the most common minerals (Bird et al., 1991; Andersen et al., 1998; Nielsen et al., 2005).

The Platinova reef is an example of a rare type of PGE deposit, typified by a Pd-Au-Cu-dominant, Ni-Pt-S poor ore mineralogy. It is generally accepted to be an orthomagmatic deposit with little post-solidification remobilisation, but the nature of the mineralisation process remains contentious (Bird et al., 1991; Andersen et al., 1998; Nielsen et al., 2005; Andersen, 2006; Holwell and Keays, 2014; Holwell et al., 2015; Nielsen et al., 2015). The current knowledge favours a multistage model of sulphide PGE-scavenging followed by sulphide dissolution and noble metal upgrading in the later stages of crystallization with separation of Fe- and Si-rich immiscible liquids playing a key role for redistribution of the precious metals (Nielsen et al., 2015).
Figure 12. (a) Two-dimensional WSW-ENE schematic cross-section of the Skaergaard intrusion showing the three main series of the intrusion. LS: Layered Series (crystallising from the floor and upward); UBS: Upper Border Series (crystallising top down) and MBS: Marginal Border Series (crystallising inwards from the walls). The LS and UBS meet at the Sandwich Horizon (SH). HZ-LZ (a, b, c), MZ and UZ (a, b, c) denote the Hidden, Lower, Middle, and Upper Zones (and subzones, respectively) of the LS and is based on appearance and disappearance of cumulus minerals indicated by plus and minus signs on the figure. Equivalent zones and subzones exist for the UBS and MBS. The Platinova Reef is within the LS located in the top of the Middle Zone (MZ). Mineral abbreviations: ap, apatite; au, augite; fer, ferrobustamite; ilm, ilmenite; mag, magnetite; ol, olivine. (b) The saucer shaped model of the Platinova Reef mineralisation showing the individual PGE and Au layers (modified after: Nielsen et al., 2015).

Figure 13. Photograph of the Triple Group, hosting the PGE-Au mineralisation in the Platinova Reef of the Skaergaard intrusion. The three characteristic leucogabbro macro-rhythmic layers (L1-L3) are highlighted. The thickness of the Triple Group is approximately 100 m (C. Tegner).
Flammefjeld (Flame Mountain) takes its name from the prominent gossan on a top of a 938 m high mountain (Figure 14), which attracted the attention of exploration geologists from Nordisk Mineselskab A/S during reconnaissance in 1970. Subsequent work indicated that it hosts a prominent Climax-type Mo porphyry mineralisation, although no drilling has been carried out to date (Geyti and Thomassen, 1983, 1984). Assays of mineralised boulders range from 0.04 to 0.45 wt. % Mo with an average of 0.17 wt. % Mo (Della Valle et al., 2010). This 39.6 Ma sub volcanic complex is situated at Amdrup Fjord in East Greenland (Enclosed map) at the south-western periphery of the Kangerlussuaq intrusion (Geyti and Thomassen, 1984; Brooks et al., 2004).

The Flammefjeld igneous complex has an oval shape of 500 m x 800 m and consists of volcanic brecciae and younger quartz-feldspar porphyries and aplites (Geyti and Thomassen, 1984). The complex is dominated by a large intrusive breccia pipe hosted in syenite, which again contains several different breccia types reflecting different levels of intrusion. These brecciae include a basal aplite intrusion breccia with metre-scale angular to subrounded fragments, intermediate breccia dykes with fine-grained quartz-K-feldspar matrix, and an upper igneous breccia containing metre-sized fragments ranging from basaltic to granitic composition (Geyti and Thomassen, 1984). The quartz-feldspar porphyries occur both as an intrusive body, confined to the intrusion breccia as concentric inward-dipping ring dykes in the periphery of the complex, and as fragments in the breccia.

The mineralised body is not exposed but has been identified within the breccia pipe in blocks up to

Figure 14. Arial photograph from SE of the Flammefjeld complex showing vivid yellow and red oxidation colours (B. Thomassen). The peak is at 938 m above sea level. The large alkaline Kangerlussuaq intrusion is seen in the background.
1 m in size that are assumed to have been torn off from a deeper porphyry system. Molybdenite is mostly found in felsic pipe brecciae where it forms typical stockwork molybdenite mineralisations. Usually, the molybdenite occurs in a dense network of veinlets associated with quartz, pyrite and minor chalcopyrite (Geyti and Thomassen, 1984). Alteration is pervasive in the Flammefjeld subvolcanic complex, and includes quartz-sericite, pyritic and argillic types. Potassic alteration is limited and restricted to biotite. A conceptual model was presented by Geyti and Thomassen (1983) who envisaged an inverted-saucer shaped ore-body with a grade of up to 0.5 wt. % MoS$_2$, a diameter of 800 m, a thickness of 200 m and which is believed to be located at about 500 m below the present exposures.

Epithermal gold- and silver-bearing quartz-carbonate veins of low-sulphidation type occur within a distance of 5 km from the Flammefjeld igneous complex with which they are associated (Geyti and Thomassen, 1984; Thomassen and Krebs, 2001; Holwell et al., 2012b). In total about 40 veins have been discovered of which most are <1 m wide and generally trending WNW-ESE. The two widest veins are the Yellow Zone and Tågegangen with widths of up to 30 m and strike extent of several hundred metres (Thomassen and Krebs, 2001). The mineralisation can, according to Thomassen and Krebs (2001), be grouped into three types: (1) pyrite-Au bearing veins; (2) galena-sphalerite-pyrite veins with Ag; and (3) polymetallic chalcopyrite-tetrahedrite-tennantite-pyrite ± sphalerite ± galena veins with Au and Ag. Float samples from the veins returned maximum 38.4 g/t Au and in-situ grab samples up to 7.5 g/t Au and 1583 g/t Ag, 7.7 wt. % Cu, 10.1 wt. % Pb and 12.2 wt. % Zn. The mineralisation also has elevated Mn, As, Sb, Mo and Bi.

MALMBJERG
MOLYBDENUM DEPOSIT
Jakob K. Keiding

The Malmbjerg porphyry-molybdenum deposit is located in the Werner Bjerg complex; East Greenland (Bearth, 1959; Schassberger and Galey, 1975; Harpøth et al., 1986). It was discovered in 1954 during systematic mapping by the Danish East Greenland Expeditions and from 1958 to present more than 30 km have been drilled and adits totalling 1329 m have been excavated. With a measured and indicated resource of 329 Mt grading 0.10 wt. % Mo (KGHM, 2015), Malmbjerg ranks as one of the world’s largest molybdenum deposits. Although access is difficult, it promises to be of exploitable value and inception of mining operation at Malmbjerg was close at the time of the world economic crisis in 2008. However, there is now renewed interest in the deposit. An exploitation license was issued for the deposit in 2009 and Malmbjerg is now (2015) licensed to KGHM.

The Malmbjerg deposit is hosted in a 25.7 Ma composite alkali granite stock that was intruded into Carboniferous to lower Permian sediments (Figure 15; Brooks et al., 2004). The dominant lithology is perthite granite with a quartz-feldspar porphyry roof phase (Figure 16). It was intruded by heterogeneous porphyritic aplite and porphyritic granites (Harpøth et al., 1986). Volumetrically minor dykes of basaltic, trachytic and lamprophyric composition later intruded the complex.

Three types of mineralisation are associated with Malmbjerg: (1) stockwork molybdenite mineralisation; (2) Mo-W-bearing greisen veins; and (3) a base metal mineralisation. The stockwork molybdenite mineralisation is predominantly hosted by the perthitic granite and the quartz-feldspar porphyry roof zone and forms an inverted bowl-like structure of approximately 700 x 700 x 150 m (Figure 16). The molybdenite
occurs in veinlets ranging in thickness from < 1 mm to 5 cm. The veinlets form a stockwork of mutually offsetting veins and consist mostly of biotite, molybdenite, quartz, fluorite, magnetite and minor siderite (Harpøth et al., 1986). The greisen mineralisation is found as up to 1 m wide flat-lying veins both in the granite stock and in the surrounding contact-metamorphosed...
sediments; they locally make up more than 10% of the volume in the porphyritic aplite where they are most abundant. The veins have a mineralogy of quartz, molybdenite, wolframite, topaz and fluorite but locally also include beryl, cassiterite, siderite, pyrite, sphalerite, chalcopyrite, bismuth and bismuthinite (Harpesth et al., 1986). The base-metal mineralisation, which cuts both the stockwork and the greisen mineralisations, is hosted by sub-vertical 30 cm thick argillised fractures. It occurs mostly distally and is minor. The base metal mineralisation consists of quartz, biotite, sphalerite, chalcopyrite, galena, pyrite and siderite or dolomite, ankerite, fluorite, sphalerite and pyrite (Harpesth et al., 1986). Silicification and hydrothermal alteration is extensive both above and below the stockwork mineralisation (Figures 15 and 16) and locally completely overprints the original rock fabrics (Schassberger and Galey, 1975).

REFERENCES

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ONLINE REFERENCES


GREENLAND 189
Deposits

- Hydrothermal fields

North Atlantic Ridge

- Ridge
- Transform fault

Scale 1:5 000 000

Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 20°W
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Iceland lies within the North-Atlantic Igneous Province (NAIP) about 63-67°N, and owes its existence to an underlying mantle plume and to the Mid-Atlantic Ridge that runs through the country (Figure 1). The trace of the mantle plume, which at least dates back from the initial opening of the North-Atlantic Ocean (63 my), runs from the eastern coast of Greenland in the west across to Faeroes and Scotland in the southeast. The crustal structure of Iceland differs somewhat from the oceanic crust. It is considerably thicker due to anomalous volcanic accumulation. The rift zones of Iceland consist of fissure swarms commonly with a volcanic complex residing in the central part (Figure 2). These complexes are the locus of anomalous volcanic and intrusive activity and frequently form caldera subsidence structures. In these volcanic complexes the rock types range from basaltic to more evolved rocks such as andesite and rhyolite, either formed through fractional crystallization or partial melting of the lower crust. There the relatively shallow magmatic heat-sources and volatiles, along with highly permeable strata lead to the formation of vigorous high-temperature systems. The lifetime of a volcanic complex along with its multiphase geothermal system ranges from 0.3 to over 1 million years, where its extinction is related to the drifting out of the volcanic zone. The average rifting velocity in Iceland is about 1 cm/y in each direction and the age of the crust increases gradually away from the rift zones to a maximum of about 15 my on either side (Saemundsson 1979, Hardarson et al. 2008). The geological map in Figure 2 depicts several formations. The volcanic zones with their postglacial lava fields are surrounded by late Pleistocene areas (< 0.8 my). Farther out are the early Pleistocene areas (0.8-2.6 my) (Hjartarson & Saemundsson 2014). Evidence of 20-30 individual glaciations have been found interbedded in the Pleistocene stratapile along with inter-glacial stages (Eiriksson 2008). The extent of the ice sheet has been traced to the outer limit of the insular shelf as indicated in Figure 1. Besides putting weight on the crust, the ice sheet changes the hydrological condition of the underlying crust, causing a variable exchange between fluid availability, permeability and heat-sources and the high-temperature systems, which may in some instances induce favorable conditions for metal deposition. No major glaciations are evidenced in the volcanic sequence in the Tertiary regions (2.6-15 my) as indicated by the dominance of sub-aerial lavas. The main erosion in Iceland is due to the repeated glaciations, and this increases away from the volcanic zones, from no erosion to 1-2 km deep erosion valleys in the older regions.

Fluid geochemistry plays an important role when considering metalliferous epithermal environment. Sulfur is common in the high-temperature systems and is considered to be the main carrier of metals. Salinity is important transport agent of metalliferous complexes. This is not the case in the Icelandic environment where meteoric water dominates in the geothermal systems with chlorine content less than 200 ppm. The geothermal systems on the outer Reykjanes Peninsula in SW-Iceland, however, have higher salinities, approaching that of seawater (Hardardottir et al. 2009).

Primary gold concentrations in Icelandic volcanic formations range from 0.5 to about 20 ppb with an average value of 3.6 ppb (Zentilli et al. 1985, Nesbitt et al. 1986, Geirsson 1993). These are higher values than the 2 ppb quoted for regular mid-oceanic ridge basalt (MORB) (Peach et al. 1990).

Iceland presents a very different geological environment in comparison with Scandinavia, Greenland, Canada, Alaska and Russia. It can, however, be viewed as a “living” example of the geological environment and processes which, in earlier geological periods, created some of the metallogenic provinces now hosting deposits which are being mined elsewhere. It also illustrates processes that are to be found within the oceanic crustal environment in several parts of the Arctic Ocean.
Figure 1. Map of the Mid-Atlantic Ridge on- and offshore Iceland (map modified from Hjartarson & Erlendsson 2016).

Figure 2. Geological map of Iceland showing Quaternary, including the volcanic zones, and Tertiary formations. Also shown are active central volcanoes and eroded, fossil central volcanoes (map modified from Hjartarson & Saemundsson 2014).
Exploration for gold in Iceland dates back to the early 20th century (Kristjansson 1929, Fridleifsson et al. 1997). The early prospectors were e.g. Björn Kristjansson, a politician and a bank manager, and Einar Benediktsson, a famous poet and entrepreneur who found a number of gold-rich localities. The most notable of these are in Mogilsa and Thormodsdalur, northeast and east of the capital Reykjavik (Figure 3). The latter, a multiple quartz-vein system, was explored by tunneling and surface excavations. The exploration stopped, apparently due to the economic recession in Europe during and following World War I. Gold exploration within the city of Reykjavik led, however, to the import of a drilling rig which (though mineable gold was not found) became very useful in drilling for geothermal water in Reykjavik (Jonasson 2006).

Renewed interest for gold exploration in Iceland developed about 1989 when the close connection between thermal activity and gold deposition, even in low salinity environments, became apparent. This led to limited reconnaissance surveys financed by private and governmental sources, later followed up by the exploration companies Malmis/Melmi and Sudurvik which made extensive, low-density reconnaissance surveys in eroded sections of the country (Franzson et al. 1992, Franzson & Fridleifsson 1993, Oliver 1993, Franzson et al. 1995, 1997).
Conditions for exploration in Iceland are good as the country is relatively barren, with good bedrock exposures, and limited sedimentary and vegetation cover. The target areas are fossil high-temperature systems within exhumed -central volcanoes. The basic theme for exploration has been the epithermal origin of gold which would imply targets limited to relatively shallow parts of the geothermal systems, i.e. above about 1000 m depth (Franzson et al. 1992, Franzson & Fridleifsson 1993). Geothermal systems in Iceland have been viewed as multiphase, extending up to 1 my in age, with thermal conditions changing according to the appearance of heat sources and/or renewed pathways for the in-/outflow of geothermal fluids. Areas around larger intrusive bodies at the base of the central volcanic complexes are also of interest for exploration, mainly for possible evidence of metalliferous magmatic volatiles. It should though be taken into consideration that magmas from oceanic crustal environment are expected to have lower volatile contents than those derived from subduction or continental environments. Base metals in high concentrations have not been found in Iceland, except at two locations in the southeast, both associated with volatiles from a rhyolitic magma source (Jankovic 1970, Franzson & Fridleifsson 1993).

The main exploration methods used are stream-sediment, rock and grab sampling. A summary of the exploration results shown in Figure 3, is mainly derived from the data collection of Melmi (e.g. Karajas 1998, Franzson et al. 2013). These show the maximum gold concentrations in individual rock or sediment samples from some of the licence areas. Many of these range from 0.2 to a little over 1 ppm Au. Four areas show more anomalous values, i.e. Viddalur-Vatnsdalur in N-Iceland with up to 32.6 ppm, Hafnarfjall in W-Iceland with samples reaching up to 4.7 ppm, Mogilsa in SW-Iceland up to 5.5 ppm and last, but not least, Thormodsdalur in SW-Iceland with values reaching up to 415 ppm. Exploration in the two last-named areas has been studied further, as described below.

**Mogilsa prospect**

The Mogilsa prospect is located on Esja Mt. some 20 km northeast of the capital Reykjavik (Figure 3). The interest in the area started in 1875 when mining of a thick calcite vein system for production of lime began (Figure 4). Later studies, around 1917, indicated the presence of gold in the veins, but efforts to establish a gold mine were aborted due to lack of belief in the analytical data.
Renewed gold exploration in the late 1980s confirmed the presence of gold in the area (Franzson et al. 1992, 2008; Franzson & Fridleifsson 1993). The host rocks are Quaternary (~2 my) and apparently part of the Kollafjordur Intrusive Complex to the south (Fridleifsson 1973). The area shows high-temperature alteration with chlorite-epidote alteration around sea level and chlorite reaching up to 400 m elevation, mainly related to a NE-SW-trending, heavily sulphidized zone. This zone contains gold enrichment which was defined by profiles using BLEG (Bulk Leach Extractable Gold) analytical methods and shown in Figure 4. The anomaly is concealed towards the northeast, where it disappears under a recent landslide. The BLEG gold values in the outer zone range from 1-10 ppb and in the inner zone from 10-380 ppb. This is concomitant with increasing sulphidation, which is most intense in the inner zone, where sampling of veins shows gold contents from 0.1 to 5.5 ppm. Breccia in the core of the vein system suggests the presence of a hydrothermal explosion breccia. Hydrothermal alteration of the rocks indicates intense chloritization at the time of gold enrichment. A fluid inclusion study shows a Th-temperature range of 200-270°C which conforms to a boiling-point curve depth of 300-500 m. An unconformity occurs some 300 m above the main anomaly, seen both as a change in strata inclination and alteration. The unconformity probably represents the surface of the geothermal system at the time of ore formation, during a state of intense boiling, a very favourable condition for gold precipitation. Figure 5 shows the conceptual model of the geothermal system and the location of the anomaly deduced from the field study. It predicts the presence of a broader, underlying geothermal reservoir narrowing upwards along a tectonic lineament. The intense sulphide zone indicates the presence of an underlying, degassing magma intrusion. The Mogilsa area lies within a very popular mountain hiking route, which may lead to public reservations regarding permits for exploration drilling and exploitation.

**Thormodsdalur prospect**

The Thormodsdalur prospect is, as yet, anomalous in Iceland in relation to gold enrichment. Its location is about 10 km from the outskirts of the capital Reykjavik (Figure 3). Initial exploration of the locality was made by a local farmer and his family and then further aided by the poet and entrepreneur Einar Benediktsson: Over 300 m of excavations and tunneling were achieved during 1908-1925 by three consecutive exploration companies. The rocks were at one stage exported to Germany, but reports of their Au content remain speculative (Fridleifsson et al. 1997). The country rock is dominantly pillow-rich hyaloclastites with subordinate sub-aerial lavas. The strata dips about 12° SE. The area may belong to the Stadalur Central Volcano (1.5-2 my) located about 6 km to the northeast (Fridleifsson 1973). The prospect area is within the chabazite-thomsonite alteration zone, indicating a low temperature environment (30-50°C) and a burial depth of 300-500 m. However, data from nearby wells to the north show an underlying propylitic alteration due to a fossil high-temperature reservoir, belonging to the Stadalur Central Volcano.

The area is densely faulted, mostly by NE-SW-trending faults parallel to the rift fractures. More northerly normal faults with a dextral strike-slip component and fracture trends are also evident. Their occurrence may be related to the structural change from a normal rift to the hybrid rift-transform environment of the Reykjanes

![Figure 5. Conceptual model of the Mogilsa geothermal system (modified from Franzson et al. 1992).](image-url)
Figure 6. Map showing the alignment of the quartz-adularia vein and the location and horizontal projection of the 41 drillholes at Thormodsdalur. Excavation north of the river shown in Figure 8 (modified from Fleming et al. 2006).

Figure 7. The Thormodsdalur prospect area looking north along the trace of the quartz-adularia vein system (photo: Hjalti Franzson ca. 2000).
Peninsula to the south. The Thormodsdalur structure belongs to the latter northerly trend and has thus a transform character. This fault has been traced for about 700 m (Figures 6 and 7).

Petrographic and XRD studies show the evolution of the vein system from a zeolite assemblage to quartz-adularia indicating progressive heating of the system and lastly to a minor calcite. The Au-enriched zone belongs to the quartz-adularia assemblage. A preliminary SEM study shows Au-grains up to 20 µm across. Temperature estimates based on mineral zonation and a limited fluid inclusion study suggest a range of 180-230°C which concurs with boiling conditions in the geothermal system at approx. 300 m depth (Franzson & Fridleifsson 1993). A review of the data suggests that the deposit is categorized as a low-sulphidization, adularia-sericite, epithermal Au-Ag type (Corbett 2004). The limited wall rock alteration suggests that this part of the geothermal system may have been relatively short-lived, but intense.

Forty-one cored holes have been drilled into the vein system, totalling nearly 3000 m. These drill holes generally extend to <100 m, many of them inclined. They indicate significant grades and thicknesses confined to two shoots along the vein structure (Fleming et al. 2006). A 450 m deep, temperature-gradient hole (HS-27, see Figure 6) was, in addition, drilled slightly to the west of the vein system, where it intersected Au-enriched veins at depths down to 450 m. The Au grades of the veins in the holes are variable, which is not surprising considering the mineral evolution discussed above: they range from <0.5 ppm to a maximum of 415 ppm (40 cm core sample in one hole).

Reykjaness geothermal system

The Reykjanes sub-aerial, high-temperature field is located at the westernmost tip of the Reykjanes Peninsula, where the Mid-Atlantic Ridge emerges on land (Figure 1). It is an extensively drilled seawater dominated geothermal field, and has a geological succession made up of submarine strata of volcanic origin which could be an analogue to oceanic crust, and has furthermore been proved to precipitate well scales that are almost identical to those found in typical black smokers (Hardardottir et al. 2009, 2010; Hardardottir 2011). It is thus an ideal location to study internal structures of a high-temperature system and reservoir characteristics for comparison with equivalent submarine hydrothermal systems.
The main geological features of the field are shown in Figure 9. The area is largely covered by sub-aerial basalt lavas erupted in postglacial time along with low level hyaloclastite ridges from the last glacial stage. Tectonic structures are related to both dilation rift and transform structures. Surface geothermal manifestations include mud pools and steam vents. Heavily altered ground is found within an area extending over about 1 km² as shown by the broken line in Figure 9.

The geothermal field is harnessed for geothermal steam to produce electricity, and for that 35 wells have been drilled into the reservoir extending from a few hundred to a maximum 3200 m depth. A study of the stratigraphy shows a dominance of pillow basalt formations in the lower part but gradually changing to a succession of tuffaceous volcanic formations of Surtseyan type above about 1000 m depth with intervening shallow water fossiliferous tuffaceous sediments. These predominate up to about 100 m depth where sub-areal lavas top the sequence. The stratigraphic cross section (Figure 10) of the shallower wells shows this character which depicts an accumulation of volcanic products from deep (pillow basalt) to shallow depths (tuffaceous sediments) which finally reaches above sea-level to produce sub-aerial lavas (Franzson et al. 2002).

Intrusions are commonly found in the succession below about 800 m depth. These are mostly fine to medium grained basalt dykes. An abundance assessment suggest they may reach up to 60% of the succession at deeper levels. These act both partly as heat source and permeability structures (Franzson et al. 2002). The increasing abundance of intrusions with depth is in line with what is expected within an oceanic crust.

Hydrothermal alteration shows a progressive intensity with depth from fresh rocks through zones characterized by smectite-zeolite, mixed layer clay, chlorite-epidote, epidote-amphibole to amphibole zone at deepest level (Tomasson & Kristmannsdottir 1972, Franzson et al. 2002, Marks et al. 2011). The temperature in the geothermal system mostly follows the boiling point.
curve down to about 1200 m depth below which it becomes more water dominated. Highest temperatures found are around 340°C at about 3 km depth. The seawater dominated geothermal fluid is mined from deep aquifers which enter the wells at high pressures and ascends to the surface. At the wellhead the fluid is decompressed resulting in steam flashing which results in

Figure 10. NE-SW cross section of the Reykjanes high-temperature field based on the early wells. The location of the cross section is shown in Figure 9 (Tomasson & Kristmannsdottir 1972, Lonker et al. 1993, Hardardottir 2002).

Figure 11. Metal and trace-element concentrations (in µMoles and nMoles) from three wells in the Reykjanes geothermal system compared to black smokers (21°N EPR, TAG and 5°S at MAR) (Hardardottir et al. 2009).
heavy pipe scaling. The analysis of these scales show high-percentage of Zn, Fe, Cu, Pb, and S, and Au concentrations up to 950 ppm and Ag up to 2.5%. The fluids at the wellhead are modified as they have boiled from 1200 m depth resulting in the precipitation of sulphides (mainly ZnS) within the wells. Sampling of deep unboiled fluid in wells has been done (Hardardottir et al. 2009, 2013) to assess the unmodified reservoir fluid composition (Figure 11). A comparison between black smoker fluids compositions and those of Reykjanes show that they are very similar in most elements. It has even been argued that due to sampling problems of true black smoker fluids, the Reykjanes samples may be better representatives of such fluids (Hardardottir et al. 2009).

The Reykjanes geothermal system has been harnessed for steam for a few years allowing a monitoring of its behavior. One of the results has shown an effective permeability barrier between the system and the surrounding groundwater systems. This is explained firstly by the rapid mineral deposition in the upflow channels forming a caprock. Secondly, at the side of the geothermal system where inflow is expected into the system, studies of wells show clear evidence of very massive anhydrite deposition in all tectonic veins, which clearly indicates a near instantaneous clogging of fracture permeability. Replenishment of fluids into the geo thermal system through tectonic fractures, at least in the upper “2 km” should therefore be considered to consist of small short term dosages rather than long term massive inflows.

SUBMARINE HYDROTHERMAL SYSTEMS

Three submarine hydrothermal systems found on the Mid-Atlantic Ridge close to Iceland are shown in Figure 1. They are located at relatively shallow waters and if at boiling condition, the temperatures would lie along the hydrologically controlled boiling point curve. Reliable data is available from these three systems; Steinaholl hydrothermal field on the Reykjanes Ridge, Grimsey hydrothermal field north of Iceland, and Kolbeinsey hydrothermal field near the southern end of the Kolbeinsey Ridge. Another feature shown on Figure 1 is the outline of the maximum extent of the Icelandic ice sheet during the last glacial period. The presence of these is evidenced by glacial debris, end-moraines and erosional features (Hjartarson & Erlendsson 2015). The latter implies that hydrothermal precipitation may have been eroded and those observed today may be limited to postglacial times. This would apply to the Kolbeinsey and Grimsey hydrothermal fields, while the Steinaholl field on the Reykjanes Ridge is found at the outer margin of the ice sheet.

Scarcity of hydrothermal manifestations on the Reykjanes Ridge south of Iceland has aroused speculation. According to German & Parson (1998) transform zones are near absent for most of the Reykjanes Ridge reflecting low incidence of venting on the ridge. One explanation for the absence of black smoker deposits may be that chemical signals from thermal plumes may be disguised by the overlying very strong ocean currents. The evidence of the very effective inflow barrier created by the anhydrite, as clearly shown in the Reykjanes geothermal system data, may also point to limited inflow into black smoker systems. Indeed this may imply that a creation of a hydrothermal system on the ocean floor may just as much depend on the availability of recharge seawater as the availability of a heat source. The link between the occurrences of hydrothermal systems on the ocean floor and the transform-rift locations further south may simply be due to these locations being the only ones where fluid recharge into a heat source is sufficient to form a hydrothermal system able to penetrate to the ocean floor. This may be the underlying reason for the scarcity of seafloor hydrothermal systems on the Reykjanes Ridge.
Steinaholl hydrothermal field, Reykjanes Ridge

The Steinaholl hydrothermal field is located at the crest of the Reykjanes Ridge at 63°06.0N and 24°31.98W (Figures 1 and 12). The area was initially discovered by fishermen who caught hot stones in trawling nets. Following an earthquake swarm in 1991 a research team, led by Icelandic research institutions, revealed the existence of a manganese deposit consisting of todorokite and birnessite (Thors et al. 1992, Olafsson et al. 1991). The Reykjanes Ridge consists at this location of en-echelon arranged fissure swarms, the Steinaholl hydrothermal field is situated in the northern sector of one of these (Figure 1). A more detailed bathymetry map of the area depicting a heavily faulted rift segment, elongated volcanic ridges and smaller circular volcanoes is shown in Figure 12. There are no obvious signs of oblique tectonics in the neighbourhood of the hydrothermal field. The seafloor is at about 250-300 m depth which is quite shallow compared to the deeper vent fields further south along the Mid-Atlantic Ridge. No rock sampling has been carried out within the area, aside from the early manganese dredging mentioned above. The main evidence of the hydrothermal activity is the formation of bubble rich plumes, which have been monitored by high-frequency echo-sounder (German et al. 1994, 1996, Palmer et al. 1995, Ernst et al. 2000). These have detected CH₄, Mn and H₂ anomalies. These bubble rich plumes are indicative of an underlying boiling hydrothermal field, which suggests temperature of 250-260°C (Ernst et al. 2000, Hannington et al. 2005).

Grimsey hydrothermal field

The Grimsey hydrothermal field is located at 66°36.36N and 17°39.24W, about 20 km ENE of the island of Grimsey (Figures 1 and 13). It is located near the northeastern edge of the northerly trending Skjalfandaflói Trough where northerly-trending faults with a westerly down throw dominate. An age estimate of the fault structures suggests that the eastern margin of the trough has been very active in Holocene times (Magnusdottir et al. 2015, Riedel et al. 2001). The hydrothermal vents are aligned in a northerly trend being ca. 1 km long and ca. 350 m wide: the shape of the field indicates its relationship to underlying northerly-trending fault structures. The extent of the bubble plume is more confined as seen in Figure 14. The boiling condition of the vent field seen by the bubble plume at a depth of about 400 m indicates

Figure 12. Bathymetry map of the area around the Steinaholl hydrothermal field, Reykjanes Ridge (multibeam data from Marine Research Institute).
temperatures of about 250°C in the uppermost part of the vent field (Hannington et al. 2001). Of the known submarine hydrothermal fields in Icelandic waters, Grimsey shows the largest amount of precipitates. The vent field is divided into three venting areas (Figure 14). The northernmost one shows isolated mounds and solitary anhydrite chimneys, the central one shows coalescing anhydrite mounds and the southern one shows apparently older, though still active mounds (Hannington et al. 2001). The topography of a typical mound is shown in Figure 15. The mineralogy of the precipitates is dominantly anhydrite and talc. Sulphides are relatively insignificant and are not seen to be precipitating in great quantity at the present time. It is, however, not possible to exclude the presence of a sulphide-rich deposit in the sub-seafloor of the hydrothermal field.

Strontium isotopic data along with water chemistry and fluid inclusion studies have shown that the anhydrite precipitation is due to shallow mixing of hydrothermal fluid and seawater (Kuhn et al. 2003).

**Kolbeinsey hydrothermal field, Kolbeinsey Ridge**

The Kolbeinsey hydrothermal field is located at 67°04.98N and 18°42.96W near the southern end of the Kolbeinsey Ridge as shown in Figures

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*Figure 13. Bathymetry map north of Grimsey Island, with the location of the Grimsey and Kolbeinsey hydrothermal fields (multibeam data from Marine Research Institute, University of Iceland, National Energy Authority and Iceland GeoSurvey).*
Figure 14. The Grimsey hydrothermal field showing depth contours, geothermal features and the area of the bubbling plume (Hannington et al. 2001).

Figure 15. Sketch showing the main characteristics of an anhydrite hydrothermal mound found at Grimsey hydrothermal field (Hannington et al. 2001).
1 and 13. It lies along the western margin of a large, northerly-striking graben structure. The hydrothermal field appears to be confined to the southern part of a small volcanic ridge which rises from 200 m up to 90 m b.s.l. Samples from the field are mostly highly altered, fragmented lava flows, with precipitates of orange-redish mud and yellow-redish iron-hydroxides staining the altered basalts and minor chimneys (Lackschewitz et al. 2006). Mineral deposition is limited, which may indicate a young age of the hydrothermal field (or the volcanic formation). The surface temperature of the manifestations is assessed from its boiling condition at 90 m depth which is equivalent to ca. 180°C (Hannington et al. 2001). Higher temperatures can be expected in the underlying system if it follows the boiling point curve. The isotopic composition of the CO₂ and He gas bubbles emitted from the hydrothermal field show a definite mantle fingerprint (Botz et al. 1999).

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CHAPTER 5

SEA-FLOOR MASSIVE SULPHIDES IN ARCTIC WATERS

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Until quite recently the mid-ocean ridges in the Arctic were some of the least explored parts of the global ridge system. However, the interest for these waters has been large, e.g. because of the ultra-slow spreading rates, their proximity to continental margins and large accumulation of sediment in parts of the ridges (hydrocarbon potential), and because of ridge hot-spot interaction and unusually shallow ridge segments in the southern parts.

Systematic mapping and exploration of the ridges to the north of Iceland started in the late 1990s and a large number of active and inactive venting sites have been documented all the way from Iceland into the so-called Eurasia Basin (Figure 1, Pedersen et al., 2010 b). The Arctic ridge system includes six major segments, which from south to north are the Kolbeinsey Ridge, Mohns Ridge, Knipovich Ridge, Molloy Ridge, Lena Trough and Gakkel Ridge.

A significant part of this 4000 km long ridge system is in Norwegian waters where the Centre for Geobiology at the University of Bergen has carried out numerous expeditions.

The southern segment of the Arctic ridge system, the Kolbeinsey Ridge and the southern Mohns Ridge, is strongly impacted by the hot-spots under Iceland and Jan Mayen, leading to elevated topography and increased volcanic activity (see Pedersen et al., 2010 b and references therein, Elkins et al., in press, see also the Iceland chapter in this volume). Further north, the magmatic activity and the crustal thickness decrease, and spreading centres and rift valleys become deeper and pronounced. Whereas the ridge is oriented at a right angle to the spreading direction at the southern and northern part of the ridge system (orthogonal spreading), the central segments of the ridge system (i.e. the Mohns and the Knipovich ridges) are characterised by oblique spreading.

Grimsey Field

The Grimsey Field was discovered in 1997 and is therefore well-studied. It is situated only about 30 km off-shore of the Grimsey Island in a sedimented graben in the Tjörnes Fracture zone (Figure 1 and Figures 13 and 14 in the Iceland chapter).

Active venting occurs over an area of c. 1 km², with shallow aquifers undergoing boiling. Boiling happens because the fluids reach temperatures up to 250°C at a water depth of only around 400 m (Hannington et al., 2001). Intense gas bubble plumes reach more than 300 m above the seafloor. The fluids come up along a fault system that is connected to a deep normal fault with a reaction zone perhaps 1-2 km beneath the seafloor (Hannington et al., 2001, Kuhn et al., 2003). The field hosts mounds 20 m across and 10 m high above the seafloor, and with 1-3 m high anhydrite and talc chimneys at the top of the mounds (Hannington et al., 2001). It is believed that sulphides precipitate at depth, which is why the fluids venting are clear and depleted in metals. Hydrocarbons in the fluids form during thermogenic breakdown of organic materials in the subsurface (Riedel et al., 2001). More information can be found in the Iceland chapter.

Kolbeinsey Field

The Kolbeinsey Field is situated just to the south of Kolbeinsey Island (which is about to disappear because of erosion) at a water depth of only 90 m (Figure 13 in the Iceland chapter). The field is related to tectonic lineaments and fissure swarms in the subsurface (Botz et al., 1999). Sediment swarms at the site varies from less than
Figure 1. Overview of the active (red) and extinct (orange) vent fields, sulphide deposits (green) and hydrothermal plumes (yellow) found along the Mid-Atlantic Ridge north of Iceland: 1) Grimsey, 2) Kolbeinsey, 3) Squid Forest, 4) Seven Sisters, 5) Soria Moria, Troll Wall, Perle & Bruse, 6) Copper Hill, Aegirs Kilde, 7) Mohns Treasure, 8) Loki’s Castle, 9) hydrothermal plume, 10) sulphide deposit, 11) sulphide deposit and hot waters, 12) and 13) hydrothermal plumes (modified from Pedersen et al., 2010 b).
100 m to 3-400 m. The hydrothermal fluids have a maximum measured temperature of 130°C, which is about 40°C below the boiling point at that depth. However, vigorous boiling is observed, suggesting that a high gas content has lowered the boiling temperature. A strong magmatic gas component is evident, both from this field and the Grimsey field (Botz et al., 1999). No significant hydrothermal deposits occur around the vents, only smectite, amorphous silica and iron-hydroxides which cement the volcanic breccias. More on this field can be found in the Iceland chapter.

Squid Forest

Squid Forest is an extinct field at the Kolbeinsey Ridge at 68°N. It was discovered in 1999 during a cruise with R/V Håkon Mosby carrying the ROV “Aglantha” (Pedersen et al., 2010 b). The field is located at 900 m water depth at the plateau of a flat-topped semi-circular volcano and consists of two sites (op.cit.). The first is in a depression on a small volcanic ridge and comprises about 30 chimney structures. These are typically 2-4 m tall, 30-50 cm wide and are situated at the top of mounds in clusters. The other site is in a sediment-covered depression close to the edge of the volcano. This is a smaller site, composed of two clusters of 8-10 up to 4 m tall chimneys, and a mound with smaller chimneys on top. Based on analyses of a sampled chimney of pyrrhotite, sphalerite, barite and amorphous silica (Nygaard et al., 2003), fluid temperatures are estimated to have been 250-300°C. (Nygaard, 2004).

Seven Sisters (Syv Søstre) Field

The field is located 170 km west of Jan Mayen at the northernmost segment of the Kolbeinsey Ridge. The central part of this segment is anomalously shallow with large, young volcanic constructions reaching up to 30 m below sea level. The Seven Sisters field is linked to a row of semi-circular, flat-topped volcanoes. Active high-temperature venting is taking place at a depth of around 150 m and has given rise to unusual hydrothermal deposits that form large mounds. The vent field was discovered by a team from the University of Bergen (UiB) in 2013 and was revisited and sampled in 2014. Detailed reports on this vent field are in preparation.

Troll Wall (Trollveggen) Field

This vent field is located at the southernmost segment of the Mohns Ridge. Like the northernmost segment of the Kolbeinsey Ridge, this part of the ridge is shallow and more magmatic productive than the ridge system further to the north. Two vent fields, Troll Wall and Soria Moria, were discovered in this area in 2005 (Pedersen et al., 2005), and a third field, Perle & Bruse, was located in the same area in 2013. This shallow and very hydrothermally active part of the ridge is termed the Jan Mayen vent field area, and the vent fields in this area are collectively referred to as the Jan Mayen vent fields.

The Troll Wall field occurs along a normal fault defining the eastern wall of the rift valley at a water depth of c. 550 m (Figure 1, Pedersen et al., 2010 b, Pedersen et al. in prep). In the field there are at least 10 major sites of venting, each comprising several 5-10 m high chimneys. The chimneys emit white smoker vent fluids with temperatures up to 270°C (op. cit., Baumberger et al. in prep). As regards base metals, the chimneys and hydrothermal deposits are Zn-dominated with sphalerite as the main sulphide mineral, and with less pyrite and chalcopyrite (Pedersen et al. 2010b, Cruz, 2016). A CO2-rich gas phase is released from the vents and bubble plumes reach high above the seafloor (Steinsland et al., in prep.). Diffuse flow through talus deposits supports mats of sulphur-oxidizing bacteria. Some hundred metres west of the high-temperature vent sites diffuse, low-temperature venting occurs along fractures and faults in the rift valley floor. Here numerous mounds of yellow-brown iron hydroxide deposits have formed (Johannessen et al., in review). The Jan Mayen fields do not support a high biomass of vent-endemic fauna. No mussel beds or alvinocarid shrimps characteristic for vent sites further south on the Mid Atlantic Ridge (MAR) are present (Schander et al., 2010).

Soria Moria Field

The Soria Moria Vent Field is also situated at the Mohns Ridge and only approximately 5 km to the south of the Troll Wall Field (Figure 1). It is found on top of a volcanic ridge at a water depth of c. 700 m (Pedersen et al., 2010 a, Pedersen et al., in prep). There are at least two venting areas.
100-200 m or more across, which are underlain by lava flows. As at the Troll Wall Field, the fluids at Soria Moria reach temperatures of 270°C and are enriched in CO₂ (up to 91.7 mmol/kg, Pedersen et al., 2010, Baumberger et al., in prep). The field contains 8-9 m tall sulphide chimneys with spalerite as the main sulphide phase, as well as irregularly shaped areas of edifices 15-20 m across and up to 10 m high, which emit clear fluids (Figure 2). The edifices are composed of barite, silica and minor pyrite, sphalerite and galena. As at the Troll Wall vent field, the base metal composition of the hydrothermal deposits is Zn-dominated: they contain significant concentrations of Ag and Au.

**Perle & Bruse**

The Perle & Bruse field is located around two kilometres east of the Troll Wall field at the flank of a large central volcano that is undergoing rift- ing. Both active and extinct vent sites are present in this area and large areas with hydrothermal deposits have been mapped using an AUV carrying a synthetic aperture sonar system (SAS) (Alden et al., 2015). The active venting occurs along rift-parallel faults, and is primarily located at two hydrothermal mounds that are separated by 300 m. Vent fluid temperatures and compositions are comparable to that of the nearby Troll Wall and Soria Moria fields. Perle & Bruse also emits a CO₂-rich gas phase that forms two distinct gas flares above the vent field. The field was discovered by UiB in 2013, and it was revisited and sampled in 2014 and 2015.

**Copper Hill**

Samples of sulphide mineralized breccias were dredged at the Mohns Ridge at 72.32°N in 2000 (Pedersen et al., 2010 b). The fault zone is situated at the NW side of the rift valley at around 900 m water depth and is probably related to an oceanic core complex. Some of the matrix-supported breccias are heavily sulphide mineralized, dominated by chalcopyrite, which have been deposited during tectonic activity (Nygaard, 2004). Based on the mineralogy of both clasts and sulphides, it is suggested that alteration and mineralization took place in the epidote zone of the hydrothermal zone (temperature 330-370°C) before the breccias gradually were exhumed during faulting (Nygaard, 2004).

**Aegirs Kilde (Vent)**

Aegirs Kilde was discovered in 2015 at the Central Mohns Ridge during an UiB-lead deep sea expedition with R/V G.O. Sars and a new Norwegian ROV-system named Aegir 6000. The vent...
field is located at around 2500 m water depth at the crest of a large axial volcanic ridge. The vent field will be revisited and sampled during an expedition in 2016.

**Mohn’s Treasure**

Massive sulphides were dredged at 73.5° N at the eastern part of the Mohns Ridge in 2002 (Pedersen et al., 2010b). More than 100 kg of hydrothermal material was brought up from 2600 m depth, mainly composed of pyrite, and as porous chimney fragments with fluid channels. No seawater anomalies were recorded, thus the material most likely comes from an extinct field.

Some 16 km to the southeast a sulphide layer was recovered in gravity cores around 1.5 m subsurface (op.cit.). The layer is located on the other side of the axial valley ridge, meaning that the sulphide must come from a different source.

**Loki’s Castle (Lokeslottet) Field**

The Loki’s Castle Vent Field was discovered in 2008 and was the first black smoker vent field visited at an ultraslow spreading ridge and in Arctic waters (Pedersen et al., 2010 a). The field is located where the Mohns Ridge turns about 80° to a more northerly trend and becomes the Knipovich Ridge (Figure 1). The water depth is about 2400 m and the field is situated near the summit of a large axial volcanic ridge. Venting occurs at the top of two hydrothermal mounds that are around 20-30 metres high and approximately 200 m across. Venting is marked by a plume that rises 3-400 m above seafloor and is characterised by anomalies in Eh, CH$_4$ and H$_2$ (Pedersen et al., 2010 b, Baumberger et al. in press). Low-temperature venting occurs at the flank of the mound and this give rise to a distinct field of barite chimneys (Eickmann et al. 2014).

Although the vent field is underlain by volcanic rocks, the influence of sediments from the nearby Bear Island sedimentary fan is observed in the fluid chemistry (e.g. very high methane and ammonia values). The Loki’s Castle field hosts a unique Arctic vent fauna (Pedersen et al. 2010a).

**Hydrothermal plumes at the Knipovich Ridge**

Pronounced anomalies in Mn, CH$_4$, and adenosine triphosphate (ATP), showing hydrothermal activity, were discovered at the southernmost part of the Knipovich Ridge in 2000 (Connelly et al., 2007, Figure 1). The source of the anomalies are probably situated to the north of the volcanic ridge, where serpentinites and other ultramafic rocks outcrop at the seafloor (op.cit.).

**Sulphides in the Lena Trough**

Massive sulphides were recovered from the Lena Trough at 81° N in 1999 (Snow et al., 2001, Figure 1). The dredge recovered 30 kg of sulphides, alteration products and hydrothermal sediments. Oxidized porous material resembling chimney pieces was also recovered. There were also fragments of ultramafic rocks (serpentinite and harzburgite), indicating an ultramafic host rock. Because the field was found without any other indications before dredging, it was named “Lucky B” (resembling the finding of the Lucky Strike Field further south on the Atlantic Ridge).

Approximately 200 km further north, at nearly 83° N, another dredge recovered massive sulphides in 2002 (Edmonds et al., 2003, Figure 1). Here the Lena Trough makes a sharp eastward bend, and transitions into the Gakkel Ridge. The sulphides represent chimney fragments. Furthermore, a camera tow above the site showed shimmering water, indicating some hydrothermal activity (op.cit.).

**Hydrothermal activity at the Gakkel Ridge**

A number of hydrothermal plumes has been detected along the 1100 km long Gakkel Ridge segment of the Arctic Ridge (Figure 1, Pedersen et al., 2010 b). No sulphides have been recovered or observed. Of the nine potential vent sites, a plume at 85° E was the most active found so far (Edmonds et al. 2003, op.cit.) with a plume 1400 m thick and centered 1000 m above the seafloor. Sampling revealed high concentrations of CH$_4$ (up to 250 nM) and H$_2$ (up to 100 nM, Upchurch et al., 2007). Bacterial mats were observed at the seafloor, supported by slightly warm and reduced fluid seeps through cracks in the fresh volcanics (Shank et al., 2007).
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Deposits
- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Size and activity
- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Scale 1:8 500 000
Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 10°E
Outline of the geology of Norway

The geology of mainland Norway is dominated by the Caledonide Orogen, which extends, in Norway, over 1500 km from Stavanger in the southwest to the northernmost part of the country (Figure 1). Understanding of the geology of the Caledonides in Scandinavia is in a period of flux following recognition, through the last decade, that a model consisting of four allochthons, Lower, Middle, Upper and Uppermost, all emplaced southeast-wards onto the Baltic Shield, must be revised (Corfu et al., 2014). The four-allochthon model assumed that the Lower and Middle Allochthons were derived from Baltica, with higher allochthons derived outboard of Baltica (from the Iapetus Ocean and related arc complexes) and the highest levels including units derived from the Laurentian side of the Iapetus Ocean (Gee et al., 2008). The recognition that nappe emplacement to the southeast was followed by periods of extensional faulting and large-scale open folding, leading to the development of extensive basement culminations, and that elements in the tectonostratigraphy have been repeated more commonly than earlier recognized, has led to awareness of the need for a more complex model (Corfu et al., 2014).

Much of the coast of southwestern Norway, from Bergen to the region of Namsos, N of Trondheim is dominated by Neo- to Meso-Proterozoic basement rocks, highly deformed and metamorphosed during the Caledonide Orogeny (Corfu et al., 2015): this area is known as the Western Gneiss Complex: it includes a classic eclogite province (Eskola, 1921), one of the first in the world in which micro-diamonds were discovered (Dobrzhinetskaya et al., 1995). The province is also host to dunites which include the world-class Åheim olivine deposit. Southernmost and southeastern Norway is dominated by Mesoproterozoic rocks and by the Permian Oslo Palaeorift. The Mesoproterozoic rocks were deformed and metamorphosed in the Sveconorwegian orogeny (ca. 1140 – 900 Ma) and now form a series of crustal blocks separated by N-S trending deformation belts (Viola et al., 2013). One of the major features in the Sveconorwegian area is the Rogaland Anorthosite Province, S of Stavanger, which includes the world-class Tellnes ilmenite deposit.

The Oslo Palaeorift contains volcanic and intrusive complexes spanning the period from Late Carboniferous to Early Triassic, emplaced into a stacked sequence of Cambro-Silurian sediments which were originally deposited in basins on top of the Sveconorwegian basement, but which were subsequently deformed during the Caledonian Orogeny, and are preserved because of the formation of the Graben.

Archaean and Palaeoproterozoic rocks of the Fennoscandian Shield are exposed west of the Caledonides in northernmost Norway and southeast of the Caledonides along the borders with Sweden, Finland and Russia (Figure 1). The Shield developed by accretion from an Archaean core with banded iron formations in its north-eastern regions with, to the west, Palaeoproterozoic granite-greenstone belts (containing gold and copper-nickel-PGE ores) and, even further west, the Mesoproterozoic Transscandinavian Igneous Belt (TIB) which is exposed in windows in the Caledonide nappe sequence along the coast of N Norway: components in the TIB include major granite, pyroxene monzonite (mangerite) and anorthosite intrusions with associated Fe-Ti deposits.
Norwegian territorial waters extend from the continental shelf, dominated by Mesozoic basins, to include several sections of the Mid-Atlantic Ridge between Iceland and Svalbard (See Chapter 6: Sea-Floor Sulphides). The Centre of Geobiology at the University of Bergen has implemented a long-term research programme to investigate geological and biological processes related to the spreading system, including black smokers and related mineralization. Loki’s Castle, a field of five active hydrothermal vents at 73°N degrees on the Mid-Atlantic Ridge at a depth of 2,352 metres was discovered in 2008 by researchers from the Centre (Pedersen et al., 2010).

Figure 1. Geological map of Norway.
The numbers refer to the deposits described in the text below.
1. The Bjørnevatn banded iron ore deposit;
2. The Nussir sediment-hosted copper-silver deposit;
3. The Ørtfjell iron deposit;
4. The Løkken volcanogenic copper-zinc deposit;
5. The Engebø eclogite-hosted rutile deposit;
6. The Nordli molybdenum deposit;
7. The Gallujavri nickel-copper-PGE deposit;
8. The Raitevarri copper-gold deposit;
History of metal mining in Norway: pre 18th Century
(Based on Boyd, R. in: Eilu et al., 2012)

Bog iron ores were exploited in Norway as early as 400 BC: 14C dating shows that twenty-eight deposits had been exploited in central Norway alone prior to 200 AD (Stenvik, 2005). The oldest record of underground mining (late 12th C) is from the Akersberg silver deposit (of Permian age) in Oslo (Nilsen & Bjørlykke, 1991). Small-scale mining of copper and silver ores began in several parts of the country in the first half of the 16thC (Falck Muus, 1924). A mining law, based on those of Saxony, was introduced in 1538-39: elements of the law are still operative – a system of claims administered by mining inspectors. The Kongsberg silver mine which exploited native silver-arsenide veins of Permian age was opened in 1623 and was in operation, with short breaks, until 1958. Long-term copper mining developed in the following thirty years, at Kvikne (1630), Røros (1644) and Løkken (1654), all massive sulphide deposits in the Caledonides in the region S of Trondheim. Mining at Røros and Løkken continued until 1977 and 1987 respectively. Røros, one of the best-preserved old mining towns in Europe, is a UNESCO World Heritage site. Norway was, for much of the period of rapid development of mining, ruled by Denmark, then a significant military power in need of a ready supply of metals for its weapons industry.

Numerous iron deposits of metasomatic origin, mainly in southernmost Norway, were mined from the 17th C but few survived to the 20th C. One exception, at Ulefoss, was mined from 1650 and finally closed in 1929, though not before it had led to the formation of Ulefos Jernverk (now part of Ulefos Holding) an iron foundry whose products today include a type of stove first produced in 1766.

Metal mining - 18th – 19th Centuries

The established copper mines were joined by another long-term operation, at Folldal, in 1748. New types of metallic ore and minerals were also exploited in addition to the established copper, silver and iron mines. Cobalt arsenide ores of Mesoproterozoic age were discovered at Modum in 1772, leading to development of the Skuterud mines and the establishment of Blaafarveværket as a royal company for the production of the dye “cobalt blue” (Figure 2). At its peak, in the 1820s and 1830s, the company was Norway’s largest industrial concern and produced 80 % of the world’s “cobalt blue”. Nickel deposits of Mesoproterozoic age were discovered in Espedalen in 1837 and mining operations commenced there in 1846 and at Ertelien in 1849: the first description of the iron-nickel sulphide, later called pentlandite, was in samples from Espedalen (Scheerreer, 1845). There followed almost 100 years of semi-continuous nickel mining in Norway, including a period in the early 1870s during which Norway was the world’s major supplier (a position supplanted by New Caledonia after the discovery there of nickel laterite ores).

Mining of pyrite from massive sulphides, for use in production of sulphuric acid, began in the mid 19th C at Vigsnes and Stord and, in 1888, at Sulitjelma (also a major copper producer) and other new deposits, as well as from the established copper mines at Røros, Løkken and Folldal (Figure 3). Molybdenum deposits were discovered in the late 1800s, including the Knaben deposit which was mined from 1885 to 1973.

Several new, large deposits of iron ore were discovered before the end of the 19th C. The Neoproterozoic Dunderland deposits (including Ørfjell), just S of the Arctic Circle near Mo i Rana, were discovered in the 18th C and the Archaean
banded iron formations at Bjørnevatom in Sydvaranger in 1865, by Tellef Dahl, the second geologist to be appointed to the Geological Survey of Norway. Foreign investment was important in prospecting and mine development in the late 1800s and early 1900s, especially in northernmost Norway. The most long-lived evidence of this is the town of Longyearbyen on Svalbard, named after the founder of the Arctic Coal Company, established in 1906, the forerunner for Store Norske Spitsbergen Kulkompani which still operates coal mines on Svalbard.

**Metal mining - 20th – 21st Centuries**

Mining commenced at three major iron ore deposits within the first decade of the new century – at Sydvaranger (1906), Fosdalen (1906) and Rødsand (Fe-V-Ti) (1910). Operations also began at Dunderland but were sporadic until 1937. The first steps, which led ultimately to Norway’s major role as a producer of titanium-oxide pigments, were taken in the early 1900s, with the establishment of a company to exploit a patent for manufacture of “titanium white” pigment from titanium dioxide. Titania A/S opened mining operations on the Neoproterozoic Storgangen deposit in Rogaland in 1916: the company was taken over by National Lead (NL), in 1927, and is now part of the NL subsidiary, Kronos. In 1957 the nearby Tellnes deposit, one of the first discovered in Norway with the aid of aeromagnetic methods, was opened. It is one of the largest ilmenite deposits in the world, currently providing >7% of world production of ilmenite (BGS, 2015). Other new ore types to be exploited in the first half of the century included the SEDEX type Zn-Pb–Cu ores at Mofjell, just S of Mo i Rana, which were mined in the period 1928–1987.

The Svalbard Treaty, signed in 1920 by 14 countries, granted Norway sovereignty of the archipelago, but gave the right to own property, including mineral rights, to nationals of all the signatory countries. 42 countries have now signed the treaty. The Norwegian company, Store Norske Spitsbergen Kulkompani, was established in 1916 and currently owns three coal mines on Svalbard. The Russian company, Trust Arktikugol, established a coal mine (still in operation) at Barentsburg in 1932. A second Russian mine, at Pyramiden, was operated by Trust Arktikugol from 1927 to 1988. Several prospecting campaigns, focused on gold and iron mineralisations, have been implemented but none of these have so far led to the definition of economically viable deposits.

Occupation of Norway during World War II had an important impact on mining in Norway due to the strategic importance of certain types of ore which were not readily obtainable to Germany.
Deposits of several types, especially of nickel and molybdenum, were exploited intensively: the known reserves at two nickel deposits, Hosanger and Flåt, which had been opened early in the century, were exhausted. The period of industrial development following World War II saw the opening of numerous new sulphide mines, in some cases based on known deposits but also on new deposits found on the basis of new geophysical exploration methods and intensive prospecting. These included the Caledonian massive sulphide deposits (year opened in parentheses) Skorovatn (Zn-Cu-pyrite) (1952), Bleikvassli (Zn-Pb–Cu) (1957), Tverrfjell (Cu-Zn) (1968), Joma (Cu-Zn) (1972) and Lergruvbakken, a new deposit in the Røros province (Cu-Zn) (1973) as well as the Bruvann Ni-Cu deposit (1988). Deposits in the Palaeoproterozoic greenstone belts included Bidjovagge (Cu-Au) (1971) and Ulveryggen (Cu) (1972). More exotic ores were mined at Søve in the Neoproterozoic Fen carbonatite where niobium was mined from 1953-65. The Finnish mining giant, Outokumpu, developed an important role in mining in Norway in the 1990s with ownership or part-ownership in operations at Løkken, Tverrfjell, Joma, Bruvann and Bidjovagge. Exhaustion of a number of deposits, erratic metal prices and competition from large, easily-mined deposits in other parts of the world led to the closure of most of the remaining sulphide deposits in the last quarter of the 20th C. Skills in metallurgy developed during Norway's long mining history, combined with ready access to hydro-electric power and imported mineral raw materials have led Norway to a dominant position in Europe in production of aluminium (29% of EU production in 2013), ferro-alloys (42.8%), refined nickel (40.1%), and silicon metal (44.7%) (data from BGS, 2015). Norway is a major producer of industrial minerals, being the world's largest producer of olivine and ground calcium carbonate and the largest in Europe of flake graphite and high-purity quartz. At the time of writing there are three metal mines in Norway – the Tellnes ilmenite mine (at 58° 20' N) and the iron mines at Rana (Ørtfjell) and Sydvaranger (Bjørnevatn). Permissions to proceed with development of a rutile mine at the Engebø deposit in West Norway and of a copper-noble metal mine at the Nussir deposit in North Norway (including the proposed plans for tailings deposition) have recently been given. Nine deposits in Norway meet the specifications for deposits considered to be large or potentially large in the FODD classification (Eilu et al., 2007; Table 1): other candidates for the classification “potentially large” have been considered but were excluded because the level of exploration in these was not considered sufficient to confirm their potential. None of the deposits qualifies for the “very large” category.

### Table 1. Large and potentially large metal deposits in Norway north of 60° N (Sources: FODD database: http://en.gtk.fi/ informationservices/databases/fodd/, Nussir ASA, Ojala (2012))

<table>
<thead>
<tr>
<th>SIZE CLASS</th>
<th>LATITUDE</th>
<th>LONGITUDE</th>
<th>MAIN METALS</th>
<th>MAIN METAL %</th>
<th>OTHER METALS</th>
<th>TONNAGE MINED (MT)</th>
<th>RESOURCE + RESERVE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bjørnevatn</td>
<td>Large</td>
<td>69.65</td>
<td>30.03</td>
<td>Fe</td>
<td>32</td>
<td>140</td>
<td>380</td>
</tr>
<tr>
<td>Nussir</td>
<td>Large</td>
<td>70.46</td>
<td>24.20</td>
<td>Cu, Ag</td>
<td>1.16/15g/t</td>
<td>Au, PGE</td>
<td>66</td>
</tr>
<tr>
<td>Ørtfjell</td>
<td>Large</td>
<td>66.41</td>
<td>14.68</td>
<td>Fe</td>
<td>34</td>
<td>28.7</td>
<td>388</td>
</tr>
<tr>
<td>Løkken</td>
<td>Large</td>
<td>63.12</td>
<td>9.70</td>
<td>Cu, Zn, Co</td>
<td>2.3/1.8/0.07</td>
<td>Ag, Au</td>
<td>24</td>
</tr>
<tr>
<td>Engebo</td>
<td>Large</td>
<td>61.49</td>
<td>5.43</td>
<td>Ti</td>
<td>2.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nordli</td>
<td>Large</td>
<td>60.48</td>
<td>11.02</td>
<td>Mo</td>
<td>0.09</td>
<td></td>
<td>210</td>
</tr>
<tr>
<td>Raitevarri</td>
<td>Potentially large</td>
<td>69.27</td>
<td>24.94</td>
<td>Cu, Au</td>
<td>Mo</td>
<td>Not determined</td>
<td></td>
</tr>
<tr>
<td>Gallujavri</td>
<td>Potentially large</td>
<td>69.63</td>
<td>25.38</td>
<td>Ni, Cu, Co, PGE</td>
<td>Au</td>
<td>Not determined</td>
<td></td>
</tr>
<tr>
<td>St. Jonsfjorden</td>
<td>Potentially large</td>
<td>78.48</td>
<td>12.89</td>
<td>Au</td>
<td></td>
<td>Not determined</td>
<td></td>
</tr>
</tbody>
</table>

Skills in metallurgy developed during Norway’s long mining history, combined with ready access to hydro-electric power and imported mineral raw materials have led Norway to a dominant position in Europe in production of aluminium (29% of EU production in 2013), ferro-alloys (42.8%), refined nickel (40.1%), and silicon metal (44.7%) (data from BGS, 2015). Norway is a major producer of industrial minerals, being the world’s largest producer of olivine and ground calcium carbonate and the largest in Europe of flake graphite and high-purity quartz. At the time of writing there are three metal mines in Norway – the Tellnes ilmenite mine (at 58° 20' N) and the iron mines at Rana (Ørtfjell) and Sydvaranger (Bjørnevatn). Permissions to proceed with development of a rutile mine at the Engebø deposit in West Norway and of a copper-noble metal mine at the Nussir deposit in North Norway (including the proposed plans for tailings deposition) have recently been given. Nine deposits in Norway meet the specifications for deposits considered to be large or potentially large in the FODD classification (Eilu et al., 2007; Table 1): other candidates for the classification “potentially large” have been considered but were excluded because the level of exploration in these was not considered sufficient to confirm their potential. None of the deposits qualifies for the “very large” category.
The Neoarchaean iron deposits in the Sør-Varanger area are banded iron formations (BIF) interpreted as exhalative sedimentary ores by Bugge (1960b). They constitute characteristic members of the supracrustal sequences of the Sørvaranger–Kola terrane, which is unconformably overlain by the Palaeoproterozoic Polmak and Pasvik Greenstone Belts, being overthrust by the Inari Terrane in the southwest.

The individual lithotectonic units were largely formed and amalgamated into their present position in the period 2.90–2.73 Ga, prior to the intrusion of late orogenic granite plutons,
for example the Neiden granite, at ca. 2.5 Ga (Dobrzhinitskaya et al., 1995; Levchenkov et al., 1995). Geochronological data presented by the latter authors indicate that most of the supracrustals and the BIF units were deposited at ca. 2.8 Ga. The supracrustal gneiss complexes hosting the BIF deposits as presented in Figure 4 are predominantly composed of pelitic, semipelitic and psammitic sediments. They generally contain only subordinate sequences of mafic to intermediate metavolcanic rocks, which are, however, important carriers of the BIF units, especially those containing Fe-rich amphibolite members (Siedlecka et al. 1985, Dobrzhinetskaya et al., 1995).

The BIF deposits vary considerably in mineralogy, type of banding (on a mm to dm scale), and size. The individual bands are mainly composed of up to three of the following minerals: magnetite, quartz, hornblende, grünerite-cummingtonite, biotite, and garnet, as well as locally diopside and hypersthene (Wiik, 1966; Siedlecka et al., 1985). They can, according to the type of mineralogical banding, be subdivided into typical oxide-facies BIF composed of alternating bands of quartz and magnetite, which may grade into silicate-facies types containing bands of magnetite and Fe-rich silicates. Although the BIF units are strongly deformed, causing tectonic dismembering and changes in the original thicknesses, they were most probably deposited at different levels in the original volcano-sedimentary sequences. The BIF ores in the Garsjo Group and in the Jarfjord Gneiss frequently comprise silicate-facies ores. They are normally 1-15 m thick and commonly constitute several-kilometre-long trains of lenses with the strike length of the individual ore bodies rarely exceeding a few hundred metres.

The Bjørnevåtn area (Figure 5) contains the largest accumulation of BIF ores in the Sør-Varanger metallogenic district. The BIF units of the presently mined ore field were first discovered by the Commissioner of Mines, Tellef Dahl, in 1865 (Dahl 1891), but exploration did not begin until 1906. Full-scale mining by A/S Sydvaranger
commenced in 1910 and continued until 1997. Open pit and subordinate underground mining gave a total production of >200 Mt of ore with about 30 % Fe (magnetic) in this period. Sydvaranger Gruve AS reopened the deposit in 2009: four of the ore bodies (Bjørnevåt, Kjellmannsåsen, Fisketind and Bjørnfjell) were being mined in 2014: probable ore reserves were 154 Mt containing 30.4 % Fe (total) (Northern Iron, 2015). Total resources (proven+indicated+ inferred) have recently been stated to be 506.6 Mt with 31 % Fe (total) (Northern Iron, 2015).

The ore bodies are part of the Bjørnevåt Group, which comprises formations predominantly composed of quartzite, conglomerate, siliceous mica schist and gneiss, as well as biotite-rich mica schist and mafic to intermediate metavolcanic rocks with major BIF units. The metavolcanic rocks contain two levels of BIF ores that occur interlayered with and separated by sequences of hornblende and quartz-biotite gneisses, as well as magnetite-rich amphibolites, locally with pillow structures. These rocks are interpreted as andesitic, dacitic and basaltic volcanic rocks, respectively (Bugge, 1978b; Siedlecka et al., 1985).

The ore zones and their wall rocks are crosscut by late orogenic granite dykes and Mesoproterozoic dolerite dykes (1300–1200 Ma, Siedlecka & Nordgulen, 1996), not shown in Figures 4 and 5. The ore bodies of the upper BIF level have been the main target for open-pit mining. The shape of the ore bodies shown in Figure 5 is the ultimate result of three phases of isoclinal to tight folding. The distribution of the ores in the northern part of the mining field is governed by a synform overturned towards the west and with a moderate plunge to the SE. The ore bodies may reach a thickness of nearly 200 m and a strike length of 1–5 km. The upper BIF usually contains 30–35 % Fe (magnetic) (>50 % magnetite) and the lower BIF 10–30 % Fe (magnetic) (Siedlecka et al., 1985). The ores consist of alternating magnetite- and quartz-dominated bands (2–10 mm thick).

Additional minerals in the BIF include hornblende, grünerite, epidote, biotite and hematite, together with traces of pyrite and chalcopyrite that are mainly confined to the magnetite bands. In some parts of the mining field, the BIF comprises nearly monomineralic bands of hornblende and grünerite. In the southernmost part of the mining field, both the amphibolitic rocks and the ore zones thin out and start to interfinger with arenitic to pelitic metasedimentary rocks locally grading into conglomeratic units with pebbles of BIF and tonalite in a magnetite-rich matrix. These features indicate transformation to shallow-water palaeoconditions.
Several sediment-hosted copper deposits are associated with Paleoproterozoic volcano-sedimentary rocks exposed in tectonic windows within the Scandinavian Caledonides in western Finnmark, North Norway. The most significant of these is the Nussir deposit in the Repparfjord Tectonic Window (RTW) with indicated and inferred resources of 66 Mt of copper ore with average grade of 1.15 % Cu and payable amounts of silver and gold (Nussir ASA, 2015; JORC estimates). Ulveryggen is another major deposit within the RTW with total resources of 7.7 Mt grading ca. 0.8 % Cu. About 3 Mt of ore with an average grade of 0.66 % Cu were mined in the period 1972–1979 from four open pits in this deposit. Nussir ASA is planning to start operating both deposits. Minor, but widespread sediment-hosted copper occurrences and mineralisations are known in neighbouring tectonic windows. Small-scale mines have been opened in several of these mineralisations, e.g. the Raipas deposit in Alta, in the late 19th century (Vokes, 1955; Vik, 1986). All of these sediment-hosted Cu-deposits are characterized by a bornite-chalcopyrite-chalcocite (±neodigenite) ore mineral paragenesis and a noticeable lack of pyrite. They exhibit textural features indicative of syn-diagenetic and epigenetic mineral precipitation and localized structural reworking. They are commonly enriched in Ag and locally have elevated Au, PGE and Co contents (Sandstad et al., 2012).

The Repparfjord Tectonic Window comprises an 8 km thick sequence of Early Paleoproterozoic volcano-sedimentary rocks (Pharaoh et al., 1983; Torgersen 2015). The sediment-hosted copper deposits are located in the nearly 3 km thick Saltvatn Group (Figure 6) - a sequence of clastic metasediments deformed and metamorphosed at greenschist facies conditions during the Palaeoproterozoic Svecofennian Orogeny. Coarse-grained arkosic sandstones and...
quartz-pebble conglomerates, interpreted as braided river deposits, are the stratigraphically lowermost exposed rocks of this group and host the Ulveryggen Cu-deposit. The sandstone coarsens upward into green polymict conglomerates and purple volcaniclastic conglomerates, interpreted as alluvial fan deposits. The Nussir Cu-deposit is hosted by 3-5 m thick argillitic dolostone and dolarenite situated within grey-purple siltstones conformably overlying the conglomerates. The sedimentary rocks are assumed to have been deposited within a rapidly subsiding basin, possibly a fault-controlled half-graben in a continental arc/back-arc setting (Torgersen, 2015).

The sedimentary sequence is conformably overlain by tholeiitic basaltic tuffs, tuffites and lavas of the Nussir Group. U–Pb dating of zircons from a mafic tuffite at the base of the Nussir Group has yielded a maximum formation age of 2073 ± 23/12 Ma (Perelló et al., 2015). A minimum age of 2069 ± 14 Ma is provided by Re–Os dating of sulphides in carbonate veins cutting the Nussir metabasalts (Torgersen et al. 2015).

Structurally, the Saltvatn Group comprises an open, km-scale NE-SW trending anticline with the Nussir Group metavolcanic rocks on its north-western limb. The metamorphic grade is up to upper greenschist facies in the north-western part of the RTW and has been dated by K-Ar on amphibole, yielding a Svecofennian age of ca. 1840 Ma (Pharaoh et al., 1982). The area is also affected by later southeast-directed thrusting during the Silurian Caledonian Orogeny (Torgersen & Viola, 2014).

The Ulveryggen deposit is situated in coarse-grained quartzitic to feldspathic sandstone and conglomerate (Figure 7) and consists of several saucer-shaped ore bodies that are ≤ 130 m wide and at least 250 m deep, and are discordant to the bedding over a length of c. 2 km (Figure 6). On a local scale, Cu-minerals occur disseminated along bedding as matrix between clastic grains and replacement of detrital feldspar and mica, as well as in quartz veinlets and along cracks. The highest copper grades seem to be confined to a more strongly sheared part of the sequence (Sandstad et al., 2007).

The Nussir deposit is located in the upper part of the Saltvatn Group that comprises interbedded grey-purple siltstone and sandstone with three, 3-5 m thick beds of argillitic dolostone, dolarenite and dolostone (Figure 8a). The two lower beds are Cu-mineralised and located in the lowermost 10s of meters of the sequence. Only the upper of these beds extends along the entire exposed length of the sequence (c. 9 km). The Cu-mineralised beds are mainly composed of white to light gray, fine-grained,
massive to bedded (<5 cm) dolarenites (Figure 8a). The contacts with the adjacent siltstones are transitional over 1–3 meters and consist of strongly foliated, dark gray, sericite-rich dolostones (Torgersen et al., 2015). The ore minerals, mainly bornite, chalcocite and chalcopyrite (± neodigenite) occur finely disseminated and commonly comprise interstitial grains and aggregates forming the matrix in the dolarenite, but are also found enriched in irregular quartz-rich veinlets and lenses (Figure 8b). The common grain size, 0.1 - 1 mm, varies according to the size of the clastic grains. Accessory sulphide minerals include covellite, wittichenite, carrolite, pyrite, sphalerite, galena and molybdenite (Sandstad, 2010; Perello et al., 2015). Au and Ag correlate well with copper whereas the distribution of PGE mineralisation deviates in relation to the main copper ore. Platinum minerals occur mainly as microscopic grains of sperrylite, both as clusters of inclusions in bornite and as dissemination in the gangue matrix (Sandstad 2010).

The formation of the sediment-hosted copper mineralisations in the RTW is not fully understood. They resemble the classical sediment-hosted Cu-deposits of the Central-African Copperbelt and Kupferschiefer in terms of host rocks, textures and mineralogy. The tectonic setting, however, differs as an active plate margin is more likely for formation of the rocks of the RTW than the intracontinental rifting setting commonly advocated for sediment-hosted Cu-deposits elsewhere. Perello et al. (2015) argue for a structurally controlled, synorogenic formation of the Cu-deposits in the RTW based on Re-Os dating of molybdenite (model ages 1761±8 Ma and 1768±4 Ma) associated the Cu-mineralisation. Although re-mobilisation during deformation of the host rock is recorded, the major Cu sulphides more commonly comprise interstitial grains and aggregates in the dolarenite matrix. This suggests an initial diagenetic formation of the major copper mineralisation.
Rana iron ores

The stratiform, banded iron ores in the Dunderlandsdal valley N of the town Mo i Rana occur in a strongly deformed unit, the Dunderland formation, comprising various schist and carbonate units with subordinate amphibolites of intrusive origin (Figure 9). This formation belongs to the Ørtfjell Group in the Ramnåli Nappe, which is part of the Uppermost Allochton of the Norwegian Caledonides (Søvegjarto et al., 1988, 1989; Gjelle et al., 1991, Marker et al., 2012). The sequence hosting the iron ores extends from Mo-sjoen in the south almost to Tromsø in the north, and is thought to have been deposited in Middle Cryogenic times, (730-800 Ma, Melezhik et al., 2014, 2015). There are several ore deposits in the northern part of the area, but none of these are economic at present.

Due to complex folding, the stratigraphy of the Dunderland formation is very uncertain. The main host-rock lithologies comprise different types of amphibolite-facies mica schists grading into calcareous schists or quartz–garnet–mica schists (Figure 9). The schists contain abundant units of dolomite and calcite marble that occur commonly in close proximity to the iron ores (Figure 10). The contact of the ores with the host rocks can be tectonic or conformable. The conformable contacts are characterised by a thin interval of carbonate-mica schist, which is commonly rich in Mn-carbonates and Mn-silicates. The iron ore shows a conformable contact with a diamicite unit in the mine area at Ørtfjellet-Kvannvatnet (Figure 11) in the western part of the Dunderland antiform. The diamicite consists of unevenly distributed, unsorted fragments mainly of fine-grained dolostones suspended in a massive, calcareous, mica-rich schist matrix. Clast sizes range from a few millimetres to a few centimetres and, rarely, up to nearly 1 m.
The iron ores occur, in general, as a series of tectonically dismembered and densely spaced segments, rarely exceeding 4 km in length (Figure 9). They occur both as linear units up to 30 m thick and as detached isoclinal folds with ore horizons doubled to tripled in thickness in the hinge zones. The ores are generally fine-grained (<1 mm) and show a banded distribution of Fe-oxides in a carbonate-bearing quartzitic to pelitic matrix (Bugge, 1948).

The ores can, in accordance with their contents of Fe, Mn and P, be separated into a number of sub-types. The two most important ores include low- and high- phosphorus ores. The former, containing 0.4–0.9 wt.% P$_2$O$_5$, is generally the most iron-rich type and comprises both specularitic haematite ores and magnetite ores that commonly constitute separate zones in the individual ore bodies. Where haematite is dominant, the gangue minerals are quartz, calcite, epidote and biotite, and where magnetite is dominant, the gangue minerals are quartz, calcite, biotite, hornblende and grunerite (Bugge 1978b). Bands of pyrrhotite up to 1 m thick are present at the margins of the ore zones, whereas minor amounts of pyrite are present in most ore types. The magnetite/haematite weight ratio of the ores currently being mined at Ørtfjellet-Kvannvatnet area is about 1:7 (Anders Bergvik, personal communication, 2014).

The second major type of iron formation is high-P magnetite (Fe–P) ore, which contains > 0.9 wt. % P$_2$O$_5$ and < 0.2 wt. % MnO. The Fe–P ore horizons are hosted, almost invariably, by calcareous mica schists, which commonly contain hornblende as an important mafic mineral. Although currently available geological information does not suggest any obvious stratigraphic separation between the different ore types, the lithostratigraphic relationships of the different segments of Fe, Fe–Mn and Fe–P iron ores in the Rana region indicate deposition at several closely spaced levels in the original stratigraphy (Melezhik et al., 2015).

The Ørtfjell deposit, by far the largest in the district, includes the Kvannvann mine, which currently produces ca. 2.1 Mt/a. Mining at the present level (Figure 12) will allow production until 2023.
Figure 11. Outcrop of diamicite close to the main Ørtfjellet-Kvannevann deposits.

Figure 12. Cross section of the Kvannevann ore body (Rana Gruber, 2015).
The ophiolite sequence at Løkken has been proved to be a fragment of a supra-subduction-zone (SSZ) ophiolite (Grenne, 1989a). It is a 1–2 km thick volcanic sequence composed of a sheeted dyke complex overlain by three volcanic members (op.cit.). The volcanic rocks comprise N-MORB to IAT basalts with lesser units of hyaloclastites, jasper beds, rhyolitic effusives and iron formations including sulphide, oxide and silicate layers (known as “vasskis”) (Grenne, 1989a). The ophiolite unit hosts the significant Løkken VMS deposit.

The Løkken Cu-Zn deposit was the largest ophiolite-hosted VMS deposit (i.e. Cyprus-type) in the world (Grenne, 1986, Grenne et al., 1999). It contained an original tonnage of about 30 Mt grading 2.3 % Cu, 1.8 % Zn, 0.02 % Pb, 16 g/t Ag and 0.2 g/t Au. About 24 Mt of the ore was mined in a period of 333 years (1654–1987): 6–7 Mt is still left in pillars and walls in the mine. Due to deformation, the ore body is tectonically disrupted into one major and several smaller bodies (Figure 13). The total length of the ore body is ca. 4 km, its average width is 150–200 m and its thickness ca. 50 m. This morphology can be ascribed to primary features such as subparallel faults at the seafloor, and an extensive, fissure-related hydrothermal vent system (Grenne & Vokes, 1990). An extensive feeder-zone system, found along the entire length of the deposit, is associated with the massive ore body (Grenne, 1989b; Figure 14). This system comprises a network of sulphide veins, from mm to some 10 cm.

The massive sulphide ore consists predominantly of pyrite with subordinate chalcopyrite and sphalerite, whereas galena, magnetite, haematite and bornite are minor components locally and fahlore is the most important accessory phase (Grenne, 1989b). Quartz is the main non-sulphide, constituting 12–14 % of the ore.
thick. The feeder zone has pyrite, chalcopyrite and quartz as its main constituents, and accessory amounts of sphalerite, magnetite and iron oxides. The total width of the zone is about 100 m. This part of the deposit was, because of its high Cu-content, the first to be mined at Løkken.

The Støren Group, E of Løkken, comprises a several km thick volcanic sequence mainly consisting of metabasalts (greenstones) of submarine origin (Grenne et al., 1999). Ribbon cherts, black shales and tuffitic to cherty metasediments are interlayered with the volcanic rocks. Geochemical data from the metabasalts mainly show MORB- but also WPB affinities (Gale & Roberts, 1974; Grenne & Lagerblad, 1985). Rhyolitic volcanic rocks are locally abundant in the northern part of the Trondheim Region, including a subvolcanic felsic intrusion dated to 495 ± 3 Ma (Roberts & Tucker, 1998). The Løkken and Støren units may be connected: the Løkken unit may, for example, have formed during subduction of parts of the Støren volcanics (the supra-subduction zone). The Støren unit contains the 19 Mt Tverrfjellet VMS deposit, which was closed in 1993 after production of 15 Mt of ore averaging 1.0 % Cu, 1.2 % Zn, 0.2 % Pb and 36 % sulphur. The deposit also contained ca. 4% magnetite, 10 g/t Ag and 0.1 g/t Au.
THE ENGBØ ECLOGITE-HOSTED RUTILE DEPOSIT

Are Korneliussen

The Engebø rutile deposit (Figure 15) is a 2.5 km long and up to 0.5 km wide E-W trending body of rutile-bearing eclogite on the northern side of Fordefjord in the Sunnfjord region of West Norway. Two types of titanium deposit occur in the region (Korneliussen et al., 2000): magmatic ilmenite-magnetite deposits associated with Mesoproterozoic mafic intrusions, and rutile-bearing Caledonian eclogitic rocks. The rutile-bearing eclogites are mafic intrusions that were transformed into eclogites during Caledonian high-pressure metamorphism at ca. 400 Ma, during the Scandian continent-continent collision stage in the orogeny (see Cuthbert et al., 2000 for an in-depth overview of eclogites in Western Norway). During the eclogitisation process, ilmenite in the protolith broke down, with the Fe entering into garnet and Ti into rutile. Hence, large volumes of ilmenite-bearing mafic rock were transformed into rutile-bearing eclogite. Other eclogite bodies in the Sunnfjord region (e.g. Orkheia) are also known to have significant contents of rutile (Korneliussen et al., 2000; Korneliussen, 2012).

Ore-type eclogite contains more than 3 wt % TiO$_2$ and the low-Ti eclogite < 3 wt % TiO$_2$, with large variations locally.

The major minerals in ore-type eclogite (Figure 16) are garnet (30-40 wt.%), clinopyroxene (omphacite; 30-40 wt.%) and amphibole (barroisite; 5-10 wt.%), whereas the most characteristic minor minerals are rutile, quartz, white mica (phengite) and pyrite; see Kleppe (2013) for detailed...
mineralogical information. More than 90 % of the titanium is situated in rutile, as indicated by the majority of samples in Figure 17, plotting close to the TiO₂-rutile 1/1-line. Retrograde alteration tends to alter rutile to ilmenite and occasionally to titanite, and in the TiO₂-rutile diagram such samples tend to plot below the 1/1-line.

There is a continuous variation in TiO₂ content from less than 0.5 wt.% to 5 wt.% (locally higher) as shown in Figure 17. For practical reasons the geological mapping of the deposit (Korneliussen et al. 1998) was based on TiO₂-content by distinguishing low-Ti eclogite with less than 3 % TiO₂ from high-Ti (ore-type) eclogite with more than 3 % TiO₂.

Although rutile-bearing eclogite rocks have been known in the Førdefjord area for a long time, geologist Hans-Peter Geis from the company Elkem was probably the first who realized the economic potential for rutile at Engebø (in 1973). In the following years Engebøfjellet was studied on a reconnaissance basis by NGU as well as by various companies. In 1995-97 an extensive project was carried out by the American titanium pigment producer DuPont and the petroleum company Conoco, at that time a DuPont subsidiary (now ConocoPhillips), in collaboration with NGU. Based on 49 drillholes (total 14,527 m) the company identified an ore resource at 380 Mt with 2.5-4.5 % rutile, sufficient to provide a fairly large amount of rutile concentrate (200.000 t/a) to the company’s downstream titanium pigment plants. The deposit was regarded to be mineable as an open-pit operation with mine waste deposition into the fjord nearby. The company, however, closed the project at the end of 1997.

In 2006 the Norwegian company Nordic Mining acquired the mineral rights to the prospect and continued its development. In agreement with the local community a compromise was developed to minimize the environmental impact, involving a two-stage mining operation: The ore is first to be mined from an open pit in the central part of the deposit, and thereafter underground. With this alternative, the resource estimate is 150 Mt of mineable rutile ore, which is much less than the DuPont estimate due to the different mining scenarios. Nordic Mining succeeded, in 2015, in achieving a permit for deposition of mineral waste at 300 m depth in the nearby Førdefjord. The deposit is now being developed further, and an annual production of 80,000 t of rutile concentrate based on the mining of 4 Mt rutile ore is expected to be a reality by 2019-2020.

The Engebø project is the first in the world to be based on extraction of rutile from hard rock such as eclogite. The main challenges relate to CaO-contamination of the rutile concentrate relative to quality criteria set by the TiO₂-pigment industry (Korneliussen et al., 2000).
The Permo-Carboniferous Oslo Palaeorift comprises two half grabens containing sedimentary rocks of Cambro-Silurian age, deformed during the Caledonian Orogeny and subsequently overlain by volcanites and truncated by numerous granitic batholith massifs. The volcanites comprise early fissure eruptives of tholeiitic and alkali basaltic composition overlain by thick sequences of trachy-andesitic rhomb porphyry lavas, which gave way to the development of central volcanoes with late-stage cauldron formation. The igneous activity ended in the Permo-Triassic with the emplacement of numerous batholiths dominated by monzonites, syenites and granites (Ihlen, 2012). The metallogeny of the rift is correspondingly varied (Ihlen, 1986), the most important types of deposits being:

- Polymetallic vein deposits along the western margin of the southern half-graben. The most important of these are the calcite veins with native silver and cobalt arsenides cross-cutting Mesoproterozoic sulphidic paragneisses at the town of Kongsberg.
- Skarn deposits of Fe, Cu, Bi, Zn, Pb, Mo and/or W at the contacts of granite intrusions with Cambro-Silurian calcareous units (marbles).
- Special-metal deposits of Nb, REE and Zr found in aphyric trachyte flows, such as that at Sæteråsen in the Vestfold Lava Plateau, in the southern half-graben. These may be of economic interest.
• Apatite-ilmenomagnetite mineralizations occur in most of the monzonitic intrusions inside the Rift. The most important deposits are in the larvikite massif at Kodal in the southernmost part of the exposed plutonic rocks in the Rift.

• Porphyry Mo deposits are the most important type of deposit in economic terms. They are associated with central granite stocks in a number of cauldrons throughout the province, and include the Nordli deposit in the Hurdal Batholith of the northern half-graben.

Comprehensive overviews of the metallogeny of the Oslo Rift can be found in Ihlen & Vokes (1978), Olerud & Ihlen (1986) and in Bjørlykke, Ihlen & Olerud (1990).

The Hurdal batholith, located in the northernmost part of the Rift, and its contact zones contain a range of types of molybdenum mineralizations (Ihlen, 2012), but by far the most important is the Nordli deposit, a stockwork porphyry mineralization discovered by Norsk Hydro in 1978 and intensively explored up to 1983.

A drilling programme led to definition of a tonnage of 200 Mt grading 0.14 % MoS$_2$ (cut-off 0.05 % MoS$_2$), which is thought to be the largest Mo deposit in Europe. Further drilling, carried out by Intex in 2006-2008, led to a minor adjustment of the reserve estimate to 210 Mt grading 0.13 % MoS$_2$ (cut-off 0.07 % MoS$_2$) (Intex Resources, 2015). Intex Resources has an exploitation licence valid to 2018, which confers exclusive rights to development of a mining operation, subject to other statutory requirements.

The deposit formed in the root zone of a deeply eroded and nested system of calderas in the northern part of the Hurdal batholith (Ihlen, 2012). The molybdenite stockworks are related to the emplacement and crystallisation of the composite Nordli alkali-granite stock, which postdates the major caldera-forming processes and which represents the final derivatives of a large alkali granite pluton in the area (Pedersen, 1986) (Figure 18). The ore zones form discs 150–250 m thick and 200–300 m across: they are stacked over a vertical distance of ca. 900 m. The zones were formed by the expulsion of molybdenum-bearing magmatic fluids generating hydraulic fractures in the apical parts of the host intrusions, which include early granophytic granite (top), intermediate aplogranite and late microgranite in the lower part of the Nordli stock.
GALLUJAVRI AND RAITEVARRI
– POTENTIALLY LARGE DEPOSITS
IN THE KARASJOK GREENSTONE BELT

Lars-Petter Nilsson, Peter M. Ihlen, Morten Often,
Jon Are Skaar, Rognvald Boyd

The Palaeoproterozoic Karasjok Greenstone Belt (KGB, Figure 19) is a north-trending continuation of the Central Lapland Greenstone Belt (CLGB) in Finland, extending parallel to the Norwegian-Finnish border on the eastern part of the Finnmark plateau. The CLGB hosts important ore bodies of several types, including iron, gold, and copper-nickel (Pohjolainen, 2012; Eilu et al., 2012), some of which also occur on the Norwegian side of the border. The KGB extends across a plateau covered by Arctic tundra at levels of 200-600 m a.s.l.: outcrop is sparse but the region attracted gold panners and prospecting companies in several periods in the 20th C. The two most important targets have been the Gallujavri Ni-Cu-PGE deposit and the Raitevarri Cu-Au deposit.

The Gallujavri ultramafic intrusion (Figure 20) is located 20 km NNW of the town of Karasjok and is hosted by easterly dipping psammitic rocks belonging to the Early Proterozoic Iddja-jav’ri Group (Siedlecka et al., 1985; Siedlecka & Roberts, 1996). The intrusion is considered to be at least 500 m thick, striking approximately N-S with a length of at least 5 km. Much of the northwestern part of the intrusion lies under the lake, Gallu’javri, with most of the exposures along the eastern bank of the lake. The intrusion was studied successively by A/S Sydvaranger (1976-82), Tertiary Minerals plc (2002-2003), Anglo American (2006-2010) and Store Norske Gull (2011-2013): the claims are now held by Nussir ASA.

The earlier investigations included ground geophysics, geological mapping, geochemical sampling and shallow drilling. Eleven shallow holes have been drilled - 8 by Sydvaranger and 3 by Tertiary Minerals, the deepest to 180 m below the surface. Outcrop and core logs indicate that the intrusion is dominated by olivine-pyroxene cumulates, at least partially recrystallized at amphibolite facies. Nilsson and Often (2005) describe different models proposed for the distribution of ultramafic components and mineralization in the body, suggesting that the base of the intrusion coincide with its western contact. Outcrops show the presence of up to ca. 4 % disseminated sulphides with grades of up to 0.42 % Ni and 0.42 % Cu in four separate areas of mineralization, one of which is exposed along a strike length of 500 m. Recorded noble metals grades range up to 2.45 g/t Pt+Pd+Au. Available data thus suggest quite high metal-insulphide tenors. Ore petrographic studies have revealed the presence of pentlandite, pyrrhotite and chalcopyrite but not discrete platinum-group minerals (Hagen and Nilsson, unpublished data).

High-resolution aeromagnetic (and radiometric) data were acquired for the whole of Finnmark county in the period 2007-2014, partly by industry but mainly by the Geological Survey of Norway (NGU). The northern part of the Karasjok Greenstone Belt was mapped in 2011 for Airborne Gravity Gradient (the first area in Norway to be covered using this technology). The availability of high quality aeromagnetic and gravity data allowed creation of a 3D interpretation of the shape of major metavolcanic units and of Gallujavri and other intrusions in the area (Skaar, 2014). The interpretation indicates that the exposed part of the Gallujav’ri intrusion is
Figure 19. Geological map of part of the Karasjok Greenstone Belt (from Skaar 2014), based on Henriksen (1986) and Nilsen (1986). The Raitevarri deposit is just to the south of the map border.
part of the northern extension of an intrusion which may be ca. 30 km long, plunging at a shallow angle to the SE where it reaches a depth of ca. 1 km.

The Raitevarri mineralization is hosted by a more than 25 km-long, sporadically exposed unit of quartz-hornblende-plagioclase-biotite gneiss (the Raitevarri Gneiss of Siedlecka & Roberts, 1996) approximately 30 km SW of the town of Karasjok. The gneiss unit contains widespread garnetiferous zones and sheets of irregularly formed amphibolites and garen-textured hornblendites, as well as biotite-chlorite-muscovite and quartz-kyanite-muscovite-pyrite schists. The latter schists are typical for areas showing ductile shearing. A/S Sydvaranger discovered the mineralization in 1969, and drilled 8 holes in the period up to 1976. ARCO (in the early 1980s) and RTZ have also held claims to the area, RTZ drilling 9 holes in 1994. The Geological Survey of Norway (NGU) carried out extensive geophysical, geochemical and geological investigations in the period 1988-90 (Dalsegg & Ihlen, 1991). The claims were held by Store Norske Gull from 2008-2014. The attraction for the prospecting industry has been the surface expression of the mineralization, said to be in the order of 10 km² (Ihlen, 2012b).

Figure 20. Sketch map of the surface expression of the Gallujavri intrusion (Nilsson & Othen, 2005). Ultramafic intrusives are shown in pale purple and gabbroic bodies in pale brown: the country rock is metapsammite.
Figure 21. Map of the Raitevarri mineralization and its environs (after Coppard, 1994).

Legend

Iddjajav'ri Group
- Amphibolites, meta-komatilites and hornblende gneisses (Bakkilvarri Fm)
- Graphite- and mica schists (Gål'libaike Fm)
- Raitevarri-type dioritic gneisses (Gål'libaike Fm)
- Imbricated sequence of quartzites, BIF, quartz-biotite gneisses, biotite-amphibole gneisses and meta-komatiites

Skuv’vanvarri Formation
- Quartzites

Jer’gul Gneiss Complex
- Banded gneisses and granitic intrusions

Fault
Basal thrust
Areas with Cu- and Fe- sulphide mineralization
Area of mutated vegetation
Outcrops
Drill hole
Antiform
The mineralisation is located in dioritic gneiss in the Gallibaike Formation, within an antiformal structure which plunges 10° – 20° to the SE (Figure 21). Dalsegg & Ihlen (1991) demonstrated that the mineralisation has three components:

- a generally weak dissemination of pyrrhotite, pyrite, chalcopyrite and/or sphalerite,
- foliation-parallel enrichment of the sulphides close to the northeastern flank of the complex and
- mineralisation along fault zones (revealed by IP measurements). At least part of the mineralisation is in contact with a unit of sulphidic graphite schist along the NE contact of the complex. 170 rock samples yielded maximum values of 0.9 g/t Au, 0.76 % Cu, and 0.24 % Zn.

The holes drilled by RTZ (Coppard, 1994) show great variation in content of sulphides: one hole contains an almost consistent sulphide content of 2–2.5 % from the bedrock surface down to a depth of 120 m. In certain cores the level is almost consistent at 0.5 % sulphide and in others the 0.5 % level includes sections up to 10 m thick with elevated sulphide contents (up to 4 %) (Coppard, 1994). The dominant sulphides are pyrrhotite, pyrite and chalcopyrite but the mineralization also includes a wide range of accessory and trace minerals including sphalerite, galena, molybdenite, native gold and lead and bismuth tellurides. Grades are stated to be low, invariably <0.8 % Cu in 1 m sections, with gold rarely exceeding 0.5 g/t (Coppard, 1994).

Store Norske Gull carried out an extensive programme of till sampling in 2009, and, in 2008-2009, a drilling programme of 28 holes totalling 3,443 m. The new data showed that the main mineralized zone had a NW-SE extent of 700 m and a width of 300 m: the drilling confirmed the relative continuity of this zone but also revealed the presence of a previously unknown mineralized body (Ojala, 2009; Jasberg et al. 2013). Eilu & Ojala (2011) suggest that the original mineralisation shows the following concentric pattern of geochemical zones from the core outwards: Au-Bi-Te, Cu(-Ag), Sn, Mo±As and Zn-Cd-Mo-Tl. They believe that the host gneiss is a metamorphosed calc-alkaline intrusive and that a range of factors indicate that the mineralisation formed as a porphyry-Cu-Au system. Ihlen et al. (1993) and Ihlen (2005), however, argue that the hornblende gneisses may represent deformed equivalents of sea-floor and sub-seafloor hydrothermally altered greenstones. The mineralisation is suggested to have formed both during the volcanic activity and during subsequent episodes of deformation (Braathen & Davidsen, 2000). This model gives a better explanation for both the location of the mineralised zones and the large strike length of the sheet-like body of hornblende gneisses.

THE GEOLOGY OF SVALBARD
Kerstin Saalmann

The Svalbard archipelago, located on the northwestern corner of the Barents Sea Shelf, was positioned north of Northeast Greenland prior to the Palaeogene opening of the Norwegian-Greenland Sea. It comprises rocks ranging in age from Proterozoic to Neogene, which can be broadly separated into pre-Devonian crystalline basement and post-Devonian cover rocks. The latter can be further subdivided into Devonian sedimentary rocks, Permo-Carboniferous sedimentary successions, Mesozoic sedimentary sequences and volcanic rocks, and Cenozoic sedimentary and volcanic rocks.

Pre-Devonian crystalline basement
The pre-Devonian crystalline rocks have traditionally been called the “Hecla Hoek” (Nordenskiöld, 1863). Modern geochronology has, however, shown that the basement units record
a complex evolution comprising a number of tec-
tono-thermal events, ranging in age from Late
Archaean to Silurian, of which the Grenvillian/
Sveconorwegian and particularly the Caledonian
orogenies show the largest imprints (Peucat et
al., 1989; Gee et al., 1995; Teben’kov et al., 2002;
Johansson et al., 2005).

The Caledonian rocks have been divided, based
on contrasts in pre-Devonian geological histo-
ry, into different provinces separated by N-S
to NNW-SSE-trending lineaments. It has been
proposed that these tectonostratigraphic do-
 mains represent independent terranes that were
amalgamated during the Caledonian orogeny
(e.g. Harland 1969; Harland 1997; Gee and Page
1994). The definition of these provinces and the
location of their boundaries are, however, debat-
ed. Harland (1997) distinguishes eastern, central
and western terranes while others advocate east-
ern, northwestern and southwestern terranes
(Gee and Page 1994; Gee and Teben’kov, 2004).
Nordaustlandet, Ny Friesland, and northwestern
Spitsbergen share a similar late-Mesoprotero-
zoic to Caledonian evolution with units in East
Greenland (Gee and Teben’kov, 2004; Johans-
son et al. 2005), whereas pre-Devonian rocks
along the west coast of Spitsbergen, south of
Kongsfjorden, record a different tectonothermal
history (Gee and Teben’kov, 2004; Labrousse
et al., 2008). Several units in the southwestern
province, including the subduction zone-related
Ordovician Vestgötabreen Complex at St. Jons-
fjorden, have, on the basis of lithology, stratigra-
phy and tectonic evolution, been correlated with
the Pearya Terrane in Arctic Canada (Ohta et al.,
1989; Harland, 1997 and ref. therein; Gee and
Teben’kov, 2004; Labrousse et al., 2008). This
correlation has, however, been questioned by
other authors (e.g. Petterson et al., 2010; Gasser
and Andresen, 2013).

The pre-Caledonian position of the terranes, and
the timing and mode of their juxtaposition, are
debated, including the existence and importance
of large-scale, strike-slip movements in the
range of many hundreds of kilometres (e.g.
Harland, 1969, 1997; Gee and Page, 1994; Gee
and Teben’kov, 2004; Labrousse et al., 2008;
Petterson et al., 2010; Gasser and Andresen,
2013).

Caledonian orogeny

The main collision between Laurentia and
Baltica-Avalonia corresponds to the Scandian
event (c. 440-400 Ma) for which there is clear
evidence in Nordaustlandet and northwestern
Spitsbergen (Harland, 1997; Johansson et al.,
2005). Intrusion of late- to post-tectonic gran-
ites, e.g. the 414 Ma Hornemannstoppen gran-
ite in northwestern Spitsbergen (Hjelle 1979)
and the 420-400 Ma cooling ages of compar-
able granite intrusions in the northwestern and
northeastern tectonic provinces (Dallmeyer et
al., 1990a) indicate the waning stages of Cale-
donian tectonism.

An earlier, Ordovician, tectono-metamorphic
event (ca. 475 Ma) is recorded in the high-pres-
sure rocks of the Vestgötabreen Complex of Mol-
alfella, S of St. Jonsfjorden (Dallmeyer et al.,
1990b; Bernard-Griffiths et al., 1993) (Eidem-
breen event, Harland, 1997). Comparable rocks
of the Richarddalen complex of Biskayahalvøya
in NW Spitsbergen (Figure 22) overlap in time.
These high-pressure metamorphic complexes
are related to subduction- zone environments
(i.a. Horsfield, 1972; Ohta et al., 1983; Hirajima
et al., 1988; Agard et al., 2005); fragments of
the complexes were later obducted during the main
collisional event.

Devonian basin development and
Svalbardian deformation

The Caledonian orogeny was followed by depo-
sition of a thick succession of Late Silurian and
Devonian, mainly continental sandstone and
shale (Old Red Sandstone) (Friend, 1961; Friend
et al. 1997). These deposits are preserved mainly
in NW Spitsbergen (Figure 22); isolated outcrops
are also found N of Kongsfjorden. Deposition
took place in a tectonically active basin char-
acterized by extensional block faulting (Manby
and Lyberis, 1992; Piepjohn, 2000 and refer-
ces therein) or sinistral strike-slip movements
(Friend et al., 1997; McCann, 2000) and was in-
terrupted by intervening episodes of shortening
(McCann, 2000). Basin inversion led to west-di-
rected folding and thrust faulting of the Old Red
Sandstone succession and locally also involved
slivers of the pre-Devonian basement (Piepjohn,
2000). This event, known as the Svalbardian
deformation (Vogt, 1928), was generally thought
Figure 22. Geological map of Svalbard (modified after Hjelle, 1993). The map also shows the location of gold prospect areas in western Spitsbergen (Store Norske Gull AS) and the location of the St. Jonsfjorden prospect area.
to be late Devonian in age. However, Piepjohn et al. (2000) provided evidence from Dickson Land that contractional deformation took place after the Late Famennian and before Late Tournaisian time. The view regarding the Svalbardian deformation as a major strike-slip event (Harland et al., 1974) has been questioned on structural and sedimentological grounds (Lamar et al., 1986; Manby and Lyberis, 1992) in favour of (sub-)orthogonal E-W to WNW-ESE shortening (Manby et al., 1994; Piepjohn 2000).

Late Palaeozoic and Mesozoic depositional environments and evolution

Svalbard was located N of Greenland after the Svalbardian deformation. Sedimentation during the Lower and Middle Carboniferous was still strongly influenced by extensional tectonics along reactivated, mainly N-S- to NW-SE-trending lineaments controlling the deposition of the infill of several narrow half-grabens and troughs (Cutbill & Challinor 1965; Gjelberg & Steel 1981; Steel & Worsley 1984) with the main rift pulse occurring in the Middle Carboniferous (Bashkirian to Moscovian) (Gjelberg & Steel, 1981; Gudlaugsson et al., 1998). Svalbard and North-Greenland belong to an extensive rift system between Greenland and Norway; rift basin systems in the southwestern Barents Sea could be connected with contemporaneous, approximately N-S-trending, fault-bounded basins in the Arctic (Gudlaugsson et al., 1998). Lower to Middle Carboniferous sediments occurred in three main N-S- to NW-SE-oriented half-graben systems, the Billefjorden Trough along the Billefjorden Fault Zone, the Inner Hornsund Trough in southern Spitsbergen, and the St. Jonsfjorden Trough in Oscar II Land between Kongsfjorden and Isfjorden (Steel & Worsley 1984; Dallmann et al., 1999a). Lower Carboniferous siliciclastic sediments containing coal beds are followed by a change from humid to arid climate in the Middle Carboniferous and deposition of shallow marine carbonate and evaporitic sequences in a restricted- to open-marine, warm-water, tropical, semi-arid environment in late Carboniferous and Permian times. Increased subsidence during late Permian times resulted in a transgression and chert sedimentation on a deeper shelf environment (Steel & Worsley 1984). The strata are characterized by spiculitic cherts with intercalated black shales and silicified limestones passing to glauconitic sandstones (Steel & Worsley 1984; Blomeier et al., 2011).

Stable platform conditions continued during the early Mesozoic (Steel & Worsley 1984; Mørk et al., 1982). Changing facies and lithology distinguish Triassic rocks from underlying Permian strata which they overly with a distinct disconformity (Harland, 1997; Mørk et al., 1999). Mesozoic sedimentation is characterized by siliciclastic coastal and deltaic progradations into a wide shelf basin. Triassic to Early Jurassic strata consist of deltaic, coastal and shallow marine deposits, followed by the establishment of deeper shelf environments during the Middle Jurassic to Early Cretaceous (Mørk et al., 1982). The dark Triassic marine shales sparked interest in phosphorite (Harland, 1997), and their high organic content make them favourable as source rocks for hydrocarbons (Leith et al. 1993; Spencer et al., 2011). Other important source rocks for hydrocarbons are Late Jurassic shales (Golonka, 2011; Spencer et al., 2011).

A return to shallow shelf, fluvial and deltaic sedimentary systems in the Early Cretaceous (Gjelberg & Steel 1995) has been attributed to processes linked to advanced rifting in the North Atlantic and Arctic Basin and the break-up of Laurentia and Europe (Worsley, 2008), finally leading to uplift of the whole northern margin in the late Cretaceous (Worsley, 2008) and, accordingly, to a lack of deposits of this age in Svalbard as well as erosion and removal of Palaeozoic-Mesozoic strata, particularly in the northern parts of the archipelago (Steel & Worsley 1984). Late Jurassic and Early Cretaceous lava flows and intrusion of dolerites, common in eastern Svalbard and Kong Karl Land, are also related to the initial break-up of Pangaea. This mafic magmatism has been linked to the High Arctic Large Igneous Province (HALIP) (Tarduno et al., 1998; Maher, 2001). The age of mafic magmatism in
Svalbard is poorly constrained because of the wide spread of Ar/Ar and K/Ar ages; more recent data may, however, suggest a shorter event at around 124 Ma (Senger et al., 2014).

Development during the Cenozoic plate tectonic reconfiguration of the Arctic

Plate tectonic reconfiguration leading to the opening of the Arctic and North Atlantic oceans during the Palaeogene caused a tectonic overprint and reactivation of pre-existing structures and formation of the NNW-trending West Spitsbergen Fold and Thrust Belt which extends over 300 km along from Sørkapp in the south to Kongsfjorden in the north (Birkenmajer, 1981; Dallmann et al., 1993). The tectonic framework strongly influenced the deposition of Palaeogene and Neogene sediments. The Palaeogene sediments comprise a coal-bearing succession of conglomerates, sandstones and shales deposed in several individual, probably isolated basins (Livšic 1974; Manum & Throndsen 1986; Atkinson, 1963; Dallmann et al., 1999b) with the Forlandsundet Graben and Central Basin being the largest. Smaller outliers of Palaeogene strata occur south of Kongsfjorden (Ny Ålesund), south of Bellsund (Renardodden), and at Sørkapp (Øyrlandet). The present geometry of the Central Basin, as exposed in southern and central Spitsbergen, is a NNW-SSE-striking asymmetric synclinorium which has a steep western limb and is separated from the underlying lower Cretaceous strata by a décollement thrust (Dallmann et al., 1993; Paech, 2001). Its Palaeocene to Upper Eocene strata are interpreted to have been deposited in a foreland basin position with respect to the West Spitsbergen Fold and Thrust Belt (Kellogg, 1975; Steel et al., 1981; Steel & Worsley, 1984; Helland-Hansen, 1990). Correlation of Palaeogene sediments within the other, smaller basins with the Central Basin succession is, however, difficult.

Palaeogene coal is being mined at Longyearbyen (Grube 7) by Store Norske Spitsbergen Grube- kompani (SNSG) and at Barentsburg (Trust Artikugol). SNSG operated mines at Svea Nord and Lunckefjell until 2015 when the mines were put on care-and-maintenance due to the low prices of coal (Store Norske Spitsbergen Kulkompani, 2015).

Opening of the Greenland-Norwegian Sea and the Eurasian Basin initiated the separation of Svalbard and Greenland by means of dextral movements along the DeGeer Fracture Zone and Hornsund Fault Zone acting as transform structures between two ridge segments (Talwani & Eldholm, 1977; Srivastava, 1985; Harland, 1969; Harland & Horsfield, 1974; Mosar et al., 2002a). The West Spitsbergen Fold and Thrust Belt formed parallel to these fracture zones and has been described as a typical transpressive belt (Harland, 1969; Lowell, 1972; Harland & Horsfield, 1974; Steel et al. 1985), but its origin is still debated and different models have been proposed (e.g. Maher & Craddock, 1988; Nøttvedt et al., 1988; Dallmann et al., 1993; Lyberis & Manby, 1993a; CASE Team, 2001; Saalmann & Thiedig, 2002). The Caledonian basement is involved in thrust belt tectonics and is predominantly exposed in the western, “thick-skinned style” part of the thrust stack. The eastern and northeastern parts of the fold belt, closer to the foreland, are characterized by a “thin-skinned” tectonic style with typical structures of fold and thrust belts (Bergh & Andresen, 1990; Welbon & Maher, 1992; Dallmann et al., 1993; Braathen & Bergh, 1995; Bergh et al., 1997; von Gosen & Piepjohn, 2001; Saalmann & Thiedig, 2002).

The Forlandsundet Graben shows a complex, multiple-stage tectonic evolution which probably commenced already during the formation of the West Spitsbergen Fold Belt and developed mainly in Eocene-Oligocene times (Lepvrier, 1990; Gabrielsen et al., 1992; Kleinspehn & Teysier, 1992). The latest tectonic developments, represented by E-W extension, are related to post-Eocene passive continental margin development to the west when Svalbard was separated from Greenland. Young magmatic activity is expressed by Miocene to Pliocene plateau basalts in northern Spitsbergen (Burov & Zagruzina, 1976; Prestvik, 1978) and Quaternary alkali basalt volcanic centres in the Bockfjorden area of northwest Spitsbergen (Gjelsvik 1963; Skjelkvåle et al. 1989).
The metallogeny of Svalbard is poorly known and detailed information about the individual deposits is mainly found in unpublished company reports written in Russian or Norwegian. These reports suggest, as does the review by Flood (1969) that the ore occurrences are dominated by epigenetic sulphide mineralisation, almost all of them on the main island, Spitsbergen. Occurrences of magnetite skarn, banded iron formation, gneisses with stratabound dissemination of Fe-sulphides and ores of orthomagmatic Ni-Cu and Fe-Ti-V occurring locally in the Pre-Devonian basement rocks are, on current knowledge, of minor importance. Well-known examples of epigenetic ores include the Zn-Pb mineralisations at St. Jonsfjorden, Kapp Mineral, Hornsund and Sinkholmen (Figure 22). The latter two, occurring in brecciated dolostones, are regarded by some prospectors as Mississippi Valley Type deposits. Few of these mineralisations are considered to have any economic potential. They comprise fracture-bound mineralisations of sphalerite and galena, locally accompanied by copper-bearing minerals and arsenopyrite in a gangue of carbonate and/or quartz. The mineralisations are concentrated in mainly pre-Devonian basement rocks exposed in the western parts of the West Spitsbergen Fold Belt (Flood, 1969; Harland, 1997 and references therein) where they are associated with breccias and deformation zones (Dallmann, 2015). The St. Jonsfjorden prospect area in western Spitsbergen (Figure 22) is known for its sulphide mineralisation, as reflected in the place name “Copper Camp” on
the southern shore of the fjord (Figure 23). This location also marks the northern margin of the Holmeslettfjella gold prospect area.

Scree geochemical surveys were carried out in the 1980s by the Geological Survey of Norway (NGU), Store Norske Spitsbergen Kulkompani AS (SNSK) and Norsk Hydro in cooperation leading to the discovery of the Svansen Au-As deposit north of Kongsfjorden and a system of quartz veins, some with high Cu- and/or Au-grades, in the Devonian immediately west of Woodfjord (Ihlen & Lindahl, 1988). The presence of gold and arsenopyrite in the Svansen deposit was confirmed by subsequent bedrock mapping and sampling.

**St. Jonsfjorden**

Store Norske Gull AS (SNG) was established as a subsidiary of SNSK in 2003 to carry out metal exploration on Svalbard and in the northern part of mainland Norway. SNG carried out a geochemical prospecting campaign in Oscar II Land and Haakon VII Land (Figure 22) and started exploratory activities for gold in 2009 when they revisited and resampled the St. Jonsfjorden area. A mineralised zone characterized by pyrite and arsenopyrite and containing up to 55 g/t gold was found in a thrust zone at Holmeslettfjella (Ojala, 2012). Exploration, including a drilling programme, continued in 2010, but all activity was terminated in the summer of 2013 when SNG was sold. The claims to the mineralized areas S of St. Jonsfjorden were, however, transferred to SNSK.

The geology of the St. Jonsfjorden area comprises Proterozoic to early Palaeozoic metamorphic basement rocks which are unconformably overlain by unmetamorphosed late Palaeozoic to Mesozoic sedimentary rocks. The overall structural architecture of the area is basically defined by the Caledonian orogeny and by Palaeogene brittle deformation related to the formation of the West Spitsbergen Fold Belt. The prospect area is located in the western part of the fold belt where the basement was actively involved in Palaeogene thrusting (Welbon & Maher, 1992; Ohta et al., 2000; Tessensohn et al., 2001; Manby & Lyberis, 2001).

The stratigraphy of the St. Jonsfjorden region is two-fold, consisting of metamorphosed pre-Devonian basement units and unmetamorphosed late Palaeozoic and early Mesozoic sedimentary rocks. The latter do not crop out in the prospect area and are therefore not described in detail: a stratigraphic overview is presented in Figure 24. Original stratigraphic relationships are difficult to reconstruct due to the polyphase tectono-metamorphic history of the basement rocks, further complicated by Palaeogene thrusting. Several, partly conflicting, stratigraphic schemes have been proposed by different working groups using inconsistent nomenclature. Many questions regarding the age and tectonostratigraphic position of distinct units are therefor still controversial and unresolved: no consensus exists on the regional tectonostratigraphy. Nonetheless, the presence of regional marker horizons aids the task of subdivision and correlation of large parts of the basement rock column. Such marker horizons include intercalated diamictites that are interpreted as glaciogenic deposits and metatillites related to the Ediacaran Marinoan glaciation (Kanat & Morris, 1988; Harland et al., 1993; Harland, 1997).

In Oscar II Land, including the St. Jonsfjorden area, the pre-Devonian rocks comprise the dominantly Neoproterozoic successions of the St. Jonsfjorden and Daudmannsodden Groups which are followed by metasedimentary rocks of the Comfortlessbreen Group (Figure 24). The latter contains glaciogenic diamictites and overlying carbonates and thus has a Neoproterozoic age. The St. Jonsfjorden Group consists of low-grade metamorphic carbonates, quartzites and phyllites with intercalated greenstones and metabasites. In the gold prospect area, calcareous sandstones and graphite-bearing marble in the lower portions of a drill core at Holmeslettfjella have been assigned to this group (Simonsen, 2012). Pelitic phyllites and carbonates of the Daudmannsodden Group do not crop out in the actual prospect zone but are exposed farther east and west (Figure 23). This also applies to the diamictite-bearing units of the Comfortlessbreen Group.

The Vestgötabreen Complex is an exotic tectonic of blueschist- to eclogite-facies metamorphic rocks thrust on top of the St. Jonsfjorden Group (Horsfield, 1972; Ohta et al., 1983; Ohta et al., 1995; Hirajima et al., 1988; Agard et al., 2005;
**Figure 24. Stratigraphic tables of the pre-Devonian crystalline basement (top) and post-Devonian cover sedimentary rocks (bottom) of the southern St. Jonsfjorden area. The latter are exposed east of the gold prospect. Basement lithologies are from Hirajima et al. (1988), Harland et al. (1993), Agard et al. (2005) and Dallmann (2015). Post-Devonian cover stratigraphy is after Dallmann (2015).**

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<tr>
<th>Group</th>
<th>Formation/unit</th>
<th>Lithology</th>
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<tr>
<td><strong>Bullbreen Group</strong></td>
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<tr>
<td>Holmesittfjella Formation</td>
<td>sandstone, shale</td>
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<td>Bullindren Formation</td>
<td>conglomerate</td>
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<td>Motalafjella Formation</td>
<td>carbonate rocks</td>
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<td>eclogite (melanorite),</td>
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<td><strong>Complex</strong></td>
<td>Lower unit (Blueschist facies)</td>
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<td></td>
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<td>carpholite-bearing phyllite,</td>
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<td>calc-schist, boudins of serpentinite,</td>
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<td>metabasalt and metacarbonate</td>
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<td><strong>Comforlessbreen</strong></td>
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<tr>
<td>Group</td>
<td>Quartzite</td>
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<td></td>
<td>Carbonate rocks</td>
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<td>phyllite, greenschist</td>
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<td>Schistose diorite</td>
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<td><strong>Dauimannsodden</strong></td>
<td>Konowfjellet Formation</td>
<td>marble, calcareous phylite</td>
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<td><strong>Group</strong></td>
<td>Sparrefjellet Formation</td>
<td>quartzite, metasandstone</td>
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<td><strong>St. Jonsfjorden</strong></td>
<td>Alkhornet Formation</td>
<td>carbonate rocks, calcareous</td>
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<td><strong>Group</strong></td>
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<td><strong>Trollheimen Volcanic Suite</strong></td>
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<td>basic metavolcanites</td>
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<td>tectonic contact</td>
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<th>Group</th>
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<td>Gipsåelen <strong>Group</strong></td>
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<td>Gipsåelen Group</td>
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<tr>
<td>Gripshuken Formation</td>
<td>dolomite, limestone</td>
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<tr>
<td>Wordiekammen Formation</td>
<td>dolomite, limestone</td>
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<tr>
<td>Tärnikanten Formation</td>
<td>multicoloured sandstone, shale</td>
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<td>Brøggerindlen Formation</td>
<td>polymict conglomerate, sandstone</td>
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<td><strong>Carboniferous</strong></td>
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Labrousse et al., 2008). The Complex consists of two structural units (Ohta et al., 1986) (Figure 24): the structurally upper unit consists of epidote-, garnet and glaucophane-bearing schists with lenses of ultramafic rocks and metabasite that have been metamorphosed at eclogite facies conditions of approximately 580-640°C and 18-24 kbar (Hirajima et al., 1988). The structurally lower unit is composed of Fe-Mg carpholite-bearing phyllites and calc-schists and containing serpentinite, metabasalt, and meta-carbonate boudins; the carpholite schists indicate ca. 15-16 kbar and 380-400 °C suggesting a subduction zone setting (Agard et al., 2005).

Gasser & Andresen (2013), based on detrital zircon patterns, suggest that the Vestgøtabreen Complex comprises mainly Neoproterozoic metasedimentary rocks. The origin of the metabasalts and gabbroic rocks is unresolved, but they show oceanic geochemical signatures (Bernard-Griffiths et al., 1993). The timing of high-pressure metamorphism is poorly constrained between 490 and 450 Ma (ca. 475 Ma U-Pb zircon lower intercept age, Bernard-Griffiths et al., 1993; 470-460 Ma 40Ar/39Ar white mica and 485-445 Ma Rb-Sr white mica and whole rock ages, Dallmeyer et al., 1990).

The Vestgøtabreen Complex is unconformably overlain by the Bullbreen Group (Ohta et al., 1995, Harland, 1997) which shows a marked contrast in metamorphic grade and comprises three formations (Figure 24). The Motalafjella Formation is composed of limestones and dolomites and is overlain by the Bulltinden Formation consisting of conglomerates that contain clasts derived from the underlying limestones as well as from the Vestgøtabreen Complex (Kanat & Morris, 1988). The Holmeslettfjella Formation on top is made up of turbiditic sandstones and slates (Kanat & Morris, 1988; Harland, 1997). In the gold prospect area, drilled graphitic schists and shales have been assigned to this formation and are interpreted as layers that have been hydrothermally altered by an organic-rich fluid (Simonsen, 2012). Conodonts from the limestones as well as fossils in some limestone clasts in the Bulltinden conglomerate indicate an early to middle Silurian age (Scrutton et al., 1976; Armstrong et al., 1986). The Vestgøtabreen Complex has been interpreted as a tectonic mélange or strongly deformed metasedimentary sequence of mainly Neoproterozoic age with interleaved basaltic rocks of unknown origin that has been subducted, metamorphosed and thrust onto the Proterozoic successions in Ordovician times (Gasser & Andresen, 2013; Bernard-Griffiths et al., 1993). The pebble content in the overlying conglomerates indicates exhumation of the high-pressure rocks during sediment deposition which is consistent with a reconstructed nearly isothermal exhumation path (Agard et al., 2005) and rapid cooling (Dallmeyer et al., 1990).

**Structural architecture**

Besides the specific subduction zone-related deformation in the Vestgøtabreen Complex, the pre-Devonian rocks in the St. Jonsfjorden area have been affected by several phases of folding. While there is general agreement on a Caledonian age of metamorphism and associated foliation development, the style, extent and degree of Palaeogene deformation in the basement related to West Spitsbergen Fold Belt are controversial: these are issues which affect the interpretation of the timing and origin of the gold mineralisation. Michalski et al. (2014), for instance, identify three fold generations (F1-F3), of which the first two are attributed to the Caledonian orogeny. Their F2 generation also includes the major NW–SE-trending and SE-plunging synform which is overturned to the NE and which can be traced from the southern shoreline of St. Jonsfjorden west of the Copper Camp to the SE along Holmeslettfjella and further towards the eastern slope of Motalafjella (Figure 24). This fold has been interpreted as a Palaeogene structure by other authors (Morris 1988; Manby & Lyberis 2001). Likewise, cleavage development in the Bullbreen Group has been attributed to east-vergent folding during Palaeogene thrusting based on the dominant deformation mechanism by pressure-solution which clearly post-dates peak metamorphism (Manby & Lyberis 2001). However, Michalski et al. (2014) recognize flattened and sheared clasts in the Bulltinden conglomerate in the hanging wall of a Palaeogene thrust fault indicating a west-directed sense of shear, opposite to movements along the Palaeogene thrust, and came to the conclusion that the zone represents a west-directed Caledonian thrust that has been reactivated during Palaeogene fold-belt formation. Caledonian deformation in the Bullbreen Group, in turn, implies the possible existence of a Silurian Caledonian event.
also in Oscar II Land, diverging from the view of solely Ordovician metamorphism in western Svalbard (e.g. Harland, 1997). A late-Caledonian event is corroborated by conodont studies in the Bullbreen Group indicating temperatures above 300°C (Conodont Alteration Index) (Armstrong et al., 1986).

The Palaeogene overprint of the basement rocks, related to development of the West Spitsbergen fold-and-thrust belt, is often difficult to identify with certainty; one distinct difference is that Palaeogene deformation was not accompanied by metamorphism. The late Palaeozoic–Mesozoic sedimentary successions in the inner parts of St. Jonsfjorden show typical structures of foreland-propagating fold-and-thrust belts with out-of-sequence and back thrusting complicating the overall picture (Welbon & Maher, 1992; Manby & Lyberis, 2001). Such sedimentary rocks are absent in the gold prospect area. However, in western Oscar II Land, mainly along the eastern border of Forlandsundet, basement thrust sheets are interleaved with slices of Permo-Carboniferous rocks (Hjelle et al., 1999; Ohta et al., 2000; Saalmann & Thiedig, 2002; Tesiensohn et al., 2001), demonstrating that the basement was actively involved in Palaeogene thrusting. These interleaved late Palaeozoic rocks show a strong internal deformation (Tesiensohn et al., 2001). The geometry in the western parts of the fold belt suggests a series of basement-rooted floor thrusts that involved the whole sequence and cut up-section into the cover rocks (Manby & Lyberis, 2001). The major thrust faults in the prospect area are interpreted as Palaeogene structures and are examples of such basement-dominated thrust sheets. A number of prominent N/NW-S/SE striking thrust zones can be mapped out, each thrust sheet containing a distinct pre-Devonian stratigraphic unit (Figure 23). A major thrust runs from Bulltinden at the St. Jonsfjorden coastline towards Motalafljella to the SE and carries the high-pressure metamorphic rocks of the Vestgötabreen Complex onto the overturned western limb of the above-mentioned synform (F2) of the Bullbreen Group. The thrust splits into two splay faults leading to further imbrication of the complex. The floor thrust of the Bullbreen Group thrust sheet separates the Holmeslettfjella Formation on the eastern synform limb from structurally lower limestones and phyllites of the Alkhornet Formation (St. Jonsfjorden Group) which, farther east, are in turn thrust onto the Lovliebreen Formation (Figure 23). Regarding the internal deformation of the basement thrust sheets, kink- to box-type folds as well as west-dipping, small-scale reverse faults in the pre-Devonian basement are spatially associated with the major thrust faults and are thus interpreted to be related to the West Spitsbergen fold-and-thrust belt (Michalski et al., 2014). The thrusts are dissected and displaced by normal faults which are partly parallel to the thrust faults (Figure 23). Such strike-parallel, late-tectonic normal faults are common features in the West Spitsbergen fold-and-thrust belt (e.g. Braathen & Bergh, 1995; Bergh et al., 1997; Maher et al. 1997; Saalmann & Thiedig, 2002).

St. Jonsfjorden Au-As mineralization
Scree and bedrock samples show a wide range of gold contents with the highest values of 55 g/t (in outcrops along the thrust NE of Skipperyggen), some boulders showing up to 25 g/t (northern Holmeslettfljella) and most gold values ranging from 0.1-0.2 g/t to values between 2 g/t and 6-7 g/t (Ojala, 2012). The map (Figure 23) reveals a clear spatial association between gold mineralisation and the Palaeogene thrust zones, particularly along the roof and floor thrusts of the thrust sheet containing the folded Bullbreen Group. The floor thrust carries mafic rocks and high-pressure schists of the Vestgötabreen Complex in the hanging wall onto the inverted limb of the Bullbreen Group syncline; the floor thrust also transports schists and phyllites of the Holmeslettfljella Formation and carbonates of the Motalafljella Formation onto calcareous phyllites of the St. Jonsfjorden Group. In addition to a structural control by thrust faults, elevated gold values also occur along a post-thrusting normal fault west of Bulltinden (Ojala, 2012) (Figure 23). The presence of elevated gold values over strike lengths of several kilometres in two, and possibly three structurally defined zones clearly qualifies the St. Jonsfjorden mineralization as “potentially large” (See Table 1). 3263 m of diamond drill core was drilled in three target areas, Holmeslettfljella, Copper Camp and Bulltinden. The holes reached the mineralised zone, but with a maximum of 1.8 g/t Au over 0.25 m did not reflect the high gold values of the surface samples and apparently did not intersect the high-grade
The Svansen Au-As mineralization is hosted by banded metasandstone and quartzite of assumed Proterozoic age. Numerous, mainly concordant quartz veins, pods and veinlets in cm- to dm-scales occur in the metasediments, but metre-wide quartz reefs are rare (Sandstad, 1989). Both the host rocks and the quartz bodies are strongly deformed and folded. The gold shows a spatial relationship to the quartz veins/reefs and also to strongly boudinaged portions of the metasediments (Sandstad & Furuhaug, 1990). The gold values of grab and channel samples vary from a few ppb to ca. 80 ppm. Gold has a strong positive correlation to As and S, and occurs mainly as free gold with grains up to 0.3 mm, although grain sizes under 50 µm are common. The wall rock alteration is generally weak with minor sericitization, chloritization and sulphidization (Sandstad & Furuhaug, 1990).

On a mesoscopic scale there is an apparent spatial relationship between hydrothermal quartz, gold and fold structures, and the mineralization shows a number of similarities to turbidite-hosted gold deposits. However, the linear extent of the mineralized zone, greater than 11 km, may indicate a relationship to major vertical shear zones.


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NORWAY


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Store Norske Spitsbergen Kulkompani AS (SNSK), http://www.snsk.no/

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Deposits

- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Size and activity

- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Scale 1:4 000 000

Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 15°E
CHAPTER 7

SWEDEN

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The bedrock of Sweden can be divided into six major lithotectonic units (Figure 1): the Svecokarelian orogen (2.0–1.8 Ga), the Blekinge-Bornholm orogen (1.5–1.4 Ga), Post-Svecokarelian magmatic and sedimentary provinces, the Sveconorwegian orogen (1.1–0.9 Ga), the Caledonian orogen (0.5–0.4 Ga) and Neoproterozoic and Phanerozoic platformal cover and igneous rocks (References on geology at the end of the chapter).

**Svecokarelian orogen**

The Svecokarelian orogen in Sweden is inferred to have formed along an active continental margin in a convergent plate boundary setting between 2.0 and 1.8 Ga. Cycles of magmatic activity and sedimentation, up to 40–50 Ma long, are a characteristic feature of the Svecokarelian orogenic development. Metamorphism under low-pressure and, in large areas, amphibolite- and even granulite facies conditions prevailed during and after crustal shortening. Large parts of the bedrock of Sweden were formed or were tectonically affected by the Svecokarelian orogeny during this time. The main lithotectonic units of the Svecokarelian orogen in Sweden are Norrbotten, Bothnia-Skellefteå and Bergslagen. These units also host the three most important mining districts in Sweden. Minor lithotectonic units include Överkalix with a bedrock similar to Norrbotten but with a different tectonic history, Ljusdal dominated by 1.87–1.84 Ga metamorphosed granitoids and Småland with rhyolites, dacites and quartz latites dated to c. 1.8 Ga and occurring as megaenclaves in felsic intrusive rocks of the Transscandinavian Igneous Belt, similar in age.

The northernmost part of the Norrbotten lithotectonic unit consists of Archaean metagranitoids, with minor gneissic rocks of supracrustal origin (Bergman et al. 2001). In Palaeoproterozoic time the Archaean rocks were rifted and intruded by ultramafic to mafic rocks, followed by deposition of sedimentary, mafic volcanic and carbonate rocks in a rift-related tectonic setting. These rocks, the Karelian supracrustal rocks (c. 2.4–1.96 Ga), make up the oldest ore-bearing formation in Sweden with iron- and copper deposits (Bergman et al. 2001). During the Svecokarelian orogeny, the older rocks were deformed and metamorphosed, and extensive igneous activity resulted in calc-alkaline andesite and a bimodal group of mafic and felsic metavolcanic rocks which, together with clastic metasedimentary rocks, were deposited on top of older rocks (Bergman et al. 2001). Several suites of intrusions in northern Sweden were formed during the orogeny.

The oldest rocks in the Bothnia-Skellefteå lithotectonic unit are turbiditic to coarse-grained sedimentary sequences with some mafic rocks. Magmatic activity during the orogeny formed the rocks of the Skellefte district, a province with both submarine and subaerial volcanic rocks deposited in volcanic arc environments. During the orogeny the supracrustal rocks were intruded by several generations of intrusive rocks of granitic to gabbroic compositions.

The oldest rocks in the Bergslagen lithotectonic unit are turbiditic metagreywackes of Palaeoproterozoic age. During the Svecokarelian orogeny intense felsic volcanism was coupled to rapid basin subsidence (Allen et al. 1996). After the waning of the volcanism the basins were filled with clastic sediments and carbonates. The main volcanic stage was followed by the intrusion of several generations of intrusive rocks of granitic to gabbroic compositions.

**Sveconorwegian orogen**

In the south-western part of Scandinavia, the bedrock shows evidence of an accretionary to terminal collisional orogenic system at 1.1–0.9 Ga. In Sweden, this orogenetic system involved subduction of continental lithosphere with the
development of eclogite and high-P granulite. A significant feature of this orogenic system in Sweden is the dominance of pre-orogenic rocks (with respect to the Sveconorwegian orogeny) including reworked basement rocks from several previous orogens.

Caledonian orogen
In Neoproterozoic to Palaeozoic time newly formed rocks, including island arcs and sedimentary rocks together with older crust, was thrust eastward on the pre-existing crust during the Caledonian orogeny (Gee et al. 2008b). The Caledonian orogen forms Sweden’s fourth most important ore district with the Köli Nappe Complex hosting massive sulphide deposits formed in an island arc setting and the Seve Nappe Complex with what is believed to be sediment-hosted copper deposits.

Blekinge-Bornholm orogen
The bedrock in the near-surface realm in the Blekinge–Bornholm orogen is dominated by magmatic rocks that formed along a 1.8 Ga ac-
tive continental margin in a convergent plate boundary setting. However, this bedrock differs from that further north, since it was affected by a younger orogenic event at 1.5–1.4 Ga, referred to as Hallandian or Danopolonian. This younger event is expressed by ductile deformation and amphibolite-facies metamorphism as well as by the intrusion of a suite of granites and syenitoids at c. 1.45 Ga.

**Post-Svecokarelian magmatic and sedimentary provinces**

Rock suites believed to be related to separate Proterozoic tectonic events and situated in the foreland to the Blekinge–Bornholm and Sveco-norwegian orogens include:

- Mesoproterozoic magmatic rocks (1.6-1.5 Ga) consisting of granite with rapakivi texture and quartz syenite, spatially associated with gabbro, anorthosite and monzodiorite in central Sweden.

These rocks are overlain by siliciclastic sedimentary rock and basalt, the latter inferred to have formed around 1.48–1.46 Ga.

- Predominantly basic dykes formed at 1.6 Ga as well as isolated intrusive bodies of nepheline syenite, probably Mesoproterozoic in age, and 1.45 Ga granite are also present in the south-eastern part of the country.

- Mesoproterozoic (1.27–1.25 Ga) dolerite sills and dykes as well as Neoproterozoic (0.98–0.95 Ga) dykes and subordinate clastic sedimentary rocks.

**Platform cover rocks**

Neoproterozoic to Phanerozoic sedimentary rocks occur as outliers at several locations in Sweden and also as basement cover rocks at the Caledonian front. In southernmost Sweden, Mesozoic and Tertiary sedimentary rocks similar to those in Denmark are found.

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**SWEDEN’S MINING HISTORY**

Sweden’s mining history goes thousands of years back in time. Early indications of iron ore mining in eastern Bergslagen come from archaeological excavations of bloomer furnaces with fragments of magnetite ore found north of the town of Uppsala. 

- "C-dating of charcoal from the furnace indicates that the mining took place during Roman Iron age (0-400 AD). It has been suggested that the ore was extracted from the nearby Stenby-Lenaberg iron deposits located a few hundred meters north of the 60 degree latitude (Kresten 1993).

Results from pollen analysis and "C-dating of charcoal from the fire setting used in ore quarrying suggest that mining at the Falun copper mine, central Bergslagen, started in early Viking time, AD 400-800 (Erikkson & Qvarfort 1996). Since then the deposit was worked continuously until the closure of the mine in 1992, with a peak in copper production in the mid-17th century. The first written document mentioning the Falu mine is from 1288 and deals with shares in the mining activity, thus making the Falu mine one of the oldest, if not the oldest, limited company in the world (Tegengren 1924, Rydberg 1979). The current incarnation of the company is the Swedish-Finnish forestry company Stora Enso. The Falu mine site was registered as a World Heritage site in 2001.

Archaeological excavations and "C-dating of the Lapphyttan blast furnace in the Norberg area indicates that iron was produced with modern techniques during the 12th century (Almevik et al. 1992).

During the end of the 16th century and the beginning of the 17th century, the demand for iron and steel increased and paved the way for technological improvements in the mines, blast furnaces and smelters in Bergslagen. From being small-scale operations, operated by local farmers, the mines, smelters and blast furnaces became larger industrial units. The industrial revolution, which took place during the late 19th century, brought steam engines, efficient pumps, railways, explosives, methods to process...
P-rich (apatite-bearing) iron ore and the use of coal instead of charcoal in the blast furnaces. These changes completely altered the industrial landscape in Bergslagen and in Sweden. The innovations, in particular the development of a Swedish railway network, made it possible to start large-scale mining of ores outside the Bergslagen district.

The iron ores in Norrbotten County, northernmost Sweden, were discovered already in the 17th century but it was not until a railway was built that they became economic. In 1888, the railway between Malmberget and the Baltic Sea coast was completed. The production at the Malmberget iron deposit then rose from 60 tons to 600,000 tons of ore annually. Four years later, the railway construction reached Narvik in Norway and opened up a link between the Baltic Sea and the Atlantic Ocean which passes several of the large iron deposits in Northern Sweden. In 1903, the production from the ores in Malmberget and Kiruna made up more than 50% of all iron ore produced in Sweden. The mines in the north have held a leading position ever since. Today most of the iron ore produced in Sweden comes from these two mines.

The improved infrastructure in northern Sweden also opened the region for exploration for other commodities. In 1930, boulders from what was to become the large Aitik Cu-deposit were found (Malmqvist & Parasnis 1972). Several years of exploration work by Boliden AB eventually led to the opening of the mine in 1968. Initial production was 2 Mt per year and after several phases of expansion, the latest approved of in 2010, plans for an annual production of 36 Mt per year by 2015 were made. This goal was achieved and surpassed already in 2014, when 39.09 Mt @ 0.2% Cu, 0.09 g/t Au and 2.14 g Ag was produced and milled (Boliden Annual Report 2014).

In 1973, a Cu-mineralised area was found 4 km west of the Kiruna iron ore deposit (Godin 1976). The initial exploration method was the recognition of the “copper plant”, Viscaria Alpina, and the mine, which was named Viscaria after the flower, was in production from 1982 to 1997. Today there are advanced plans to re-open the mine.

During the first decades of the 20th century, several small holding companies called “emissionsbolag” were created by Swedish banks. One of these was Centralgruppens Emissionsbolag with the mission to acquire stocks in new mining companies and to develop mines. Thus, it was a kind of early junior exploration company. In 1924, at a time when the company was nearly bankrupt, the Boliden Au-Cu-As deposit was found. It was put into production two years later. During the following years several new massive sulphide deposits were found west of Boliden. Today these deposits, and their host rocks, form the Skellefte District, one of the most important ore-districts in Sweden with six producing mines. Discoveries are still being made here. For example, the Björkdal Au deposit was found using geochemical methods (till-sampling) by Terra Mining AB in 1985 and went into production in 1988. It is still in production by 2016 but with a new owner. The Åkerberg Au deposit was found in 1988 by Boliden Mineral AB and was in production from...
One of the most Cu- and Zn-rich deposits ever found in the district, the Storliden deposit, was discovered in 1997 by North Atlantic Natural Resources (NAN) and was in production from 2002 to 2008. With the opening of the Kankberg Au-Te deposit in 2012 a new commodity, tellurium, was added to the list of metals produced from the Skellefte district.

The Caledonian nappe units constitute another important mineralised region whose metal potential has been known for a long time. Although the lack of infrastructure and, more recently, environmental protection policies, have limited exploitation, several mining attempts have nevertheless been made over the centuries.

One of the oldest mining operations took place at the Nasafjäll Ag deposit close to the Norwegian border (Bromé 1923, Du Rietz 1949). The deposit was mined for a few years in the mid 17th century but was never profitable. The Fröå Cu deposit was found in the mid-18th century and was mined intermittently during several periods, the latest from 1910 to 1919. Ore production reached a peak in 1917-1918, but less than 100 000 tons of ore was produced during the mine’s lifetime (Helfrich 1967). Several significant deposits in the Caledonides were found and put into production during the 1940s. Galena-bearing sandstone boulders found in 1938 led to the discovery of the Laisvall Pb-Zn deposit. Exploitation started in 1943, mainly as a measure to secure domestic supplies of Pb during World War II (Rickard et al. 1979). It turned out however, to be economic in peace time as well, and mining lasted until 2001. Exploration for sandstone-hosted Pb-Zn led to several discoveries of similar mineralisations along the Caledonian front, and the Vassbo and Guttusjö deposits 500 km to the south-southwest of Laisvall have also been in production. Exploration by the Geological Survey of Sweden in the 1970s led to the discovery of several massive sulphide deposits in the Caledonides (Zachrisson 1969). However, the only deposit that has been in production is Stekenjokk which was mined for Cu and Zn between 1976 and 1988.

The most recent mining district to be recognised in Sweden is the so-called “Gold Line” in Västerbotten County. The Gold Line refers to a southeast-trending Au anomaly detected by the State Mining Property Commission (NSG) during a till geochemistry survey in the late 1980s. The anomaly attracted exploration to this new district, and the first deposit to be found, and later put into production by Dragon Mining Ltd., was the Svartliden Au deposit. Several other Au deposits in the area are under development.

Parallel to, and in interaction with the expansion of the exploration and mining industry, there was a tremendous development of the Swedish engineering industry supplying exploration, mining and mineral processing equipment during the late 19th and 20th centuries. In 1893, ASEA built Sweden’s first three-phase electrical power transmission system for the Grängesberg iron mine. About a hundred years later, in 1987, ASEA merged with the Swiss company Brown Boveri to form ABB, a global leader in power and automation technologies. Atlas Copco was established in 1873 with the objective to manufacture and sell equipment for railway construction and operation. At the turn of the century, compressed air machinery and later pneumatic rock-drill equipment became important export products for the company. Today tools from Atlas Copco are found in mines around the world. Sandvik was founded in 1862 and from the very start, the company delivered rock-drilling equipment to the exploration and mining industries. Through mergers and acquisitions the Swedish and Finnish engineering industries have amalgamated to form multinational companies that today are market leaders as suppliers to the world’s exploration and mining industries.
The Bergslagen mining district in south-central Sweden makes up the southwestern part of the Palaeoproterozoic Svecokarelian orogen in the Fennoscandian Shield (Stephens et al. 2009). It forms one of Sweden’s most important provinces for the exploitation of metallic mineral resources. More than 7000 iron deposits, 1500 base metal deposits, 150 special metal deposits (mainly tungsten deposits) but only a few precious metal deposits are known from the area (Figures 3 - 5). In this metal-rich province, there has been continuous mining for more than 1000 years, perhaps close to 2000 years.

Defined genetic deposit types and subtypes from the area (Stephens et al. 2009, Allen et al. 1996a) include:

- Skarn iron ore (magnetite ore hosted by skarn-altered carbonates and silicates)
  - Mn-poor skarn iron ore
  - Mn-rich skarn iron ore
- Apatite-iron ore (magnetite-hematite associated by apatite and hosted by felsic volcanic rocks)
- Banded iron formations (variably quartz-banded iron oxides, locally with other silicatebands)
- Sulphide ore (base metals, locally associated with skarn iron ore)
  - Stratiform, ash-siltstone associated Zn-Pb-Ag (Âmmeberg-type)
  - Stratabound, podiform, skarn-limestone associated Zn-Pb-Ag(-Cu-Au) (Falun type)
- Manganese ore (Mn oxides)
  - Stratiform Mn-oxide analogue to skarn iron ore
  - Breccia-hosted Mn-oxides
- Special and precious metals
  - Tungsten ( scheelite in skarn and wolframite and scheelite in quartz veins)
  - REE in skarn iron ore and apatite iron ore
  - Molybdenite, granite-hosted porphyry-style or hosted by skarn

To the west and south-west, the Bergslagen district is bounded by younger batholiths, to the north by major northwest-trending, crustal-scale shear zones and to the east by the Baltic Sea (Stephens et al. 2009). It may possibly be correlated across the Baltic Sea to southwestern Finland where similar rocks and deposits can be found (Lundström & Papunen 1986). Most of the metal deposits occur in felsic metavolcanic rocks, associated with carbonate rocks and calc-silicate (skarn) rocks 1.9 to 1.8 Ga in age (Allen et al. 1996a). In a few places, the basement to the metavolcanic rocks, consisting of turbiditic metagreywacke grading upward into quartzitic rocks, is exposed (Lundström et al. 1998). A sequence of clastic metasedimentary rocks, including metargillite, quartzite and metaconglomerate lies stratigraphically above the volcanic rocks.

In most places the supracrustal rocks have been intruded by several suites of igneous rocks of which the oldest, a granitoid-dioritic-gabbroic suite, is broadly contemporaneous with the metavolcanic rocks. The supracrustal rocks and the older intrusive suite were affected by Svecokarelian deformation and metamorphism (Stephens et al. 2009).

The northern, and most mineralised part of Bergslagen, is located to the north of 60º latitude and is thus included on the Circum Arctic mineral map. Nearly half of all Swedish deposits registered in the national mineral deposit databases are located in the northern Bergslagen area.

The total amount of produced metals and metals in reserves and resources in the area is 0.49 Mt Cu, 5.7 Mt Zn, 2.2 Mt Pb, 358 Mt Fe, 46 t Au and 5185 t Ag; the precious metals are largely considered as by-products in most deposits.
<table>
<thead>
<tr>
<th>Deposit, Mine field</th>
<th>When_mined</th>
<th>Resources (Mt)</th>
<th>Mined (Mt)</th>
<th>Cu (%)</th>
<th>Zn (%)</th>
<th>Pb (%)</th>
<th>Au (g/t)</th>
<th>Ag (g/t)</th>
<th>Ref.</th>
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<td>Garpenbergsfältet</td>
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<td>1.7</td>
<td>0.37</td>
<td>115</td>
<td>2</td>
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<td>5.7</td>
<td>7.1</td>
<td>2.2</td>
<td>0.40</td>
<td>3.00</td>
<td>18</td>
<td>1</td>
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<tr>
<td>Falu gruva</td>
<td>c. 1200-1992</td>
<td>11.39*</td>
<td>3.0</td>
<td>4.0</td>
<td>1.5</td>
<td>3.00</td>
<td>18</td>
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<td>Ryllshyttlegruvan</td>
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<td>2.0</td>
<td>0.0</td>
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<td>100</td>
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<td>4.5</td>
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<td>1</td>
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<td>20.0</td>
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<td>1</td>
<td>1</td>
<td>1</td>
<td></td>
<td></td>
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<td>2.6</td>
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<td>0.36</td>
<td>18</td>
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<td>25.0</td>
<td>300</td>
<td>1</td>
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</tr>
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</table>

Table 1. Overview of the most important sulphide deposits in the Bergslagen district.

1. Official Statistics of Sweden, Metal and Mining industries. 2. Boliden Annual report 2014. 3. Tegengren 1924. 4. Press release 18 April 2012, www.kopparbergmineral. 5. Gränggruvan-gruvkarta, Bergsstatens arkiv Falun. 6. www.wikingmineral.se. 7. www.dannemoramineral.se. (*)The mined tonnage only refer to what is reported in official statistics (1833 to present), a more correct figure for the total production is in the order of 25 Mt.

Figure 3. Simplified geological map showing the extent of Palaeoproterozoic supracrustal rocks in Bergslagen. Blue-grey: metasedimentary rocks, yellow: metavolcanic rocks, deep blue: marble. Red circles show location of large deposits included on the Circum Arctic map, blue dots show medium- and smaller size iron deposit, red dots medium- and small size base metal deposits and green dots show small other metal deposits (mainly tungsten deposits). The supracrustal rocks and the metavolcanic rocks in particular, define the Bergslagen metallogenetic area. Ore deposit data from the Fennoscandian Ore Deposit Database (Eilu et al. 2010), geology modified from SGUs digital Bedrock Sweden 1:1 M map.
<table>
<thead>
<tr>
<th>Deposit. Mine field</th>
<th>When mined</th>
<th>Resources (Mt)</th>
<th>Mined (Mt)</th>
<th>Fe (%)</th>
<th>Mn (%)</th>
<th>P (%)</th>
<th>S (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grängesberg mining district</td>
<td>1783-1989</td>
<td>148</td>
<td>133</td>
<td>44.5</td>
<td>1.36</td>
<td>0.01</td>
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<tr>
<td>Dannemorafältet</td>
<td>c. 1200-1992, 2012-2014</td>
<td>62</td>
<td>32</td>
<td>38.6</td>
<td>1.98</td>
<td>0.00</td>
<td>0.21</td>
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<td>70</td>
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<td>40.0</td>
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<td>Norrberg-Morberg-Kallmorbergs-fälten</td>
<td>1858-1902</td>
<td>43</td>
<td>19</td>
<td>42.4</td>
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<td>Dannemorafältet c. 1200-1992, 2012-2014</td>
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<tr>
<td>Håksbergsfältet</td>
<td>1858-1979</td>
<td>37</td>
<td>20</td>
<td>35.4</td>
<td>0.18</td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>Blötbergsfältet</td>
<td>1859-1979</td>
<td>40</td>
<td>16</td>
<td>40.0</td>
<td>0.16</td>
<td>1.45</td>
<td>0.02</td>
</tr>
<tr>
<td>Risbergsfältet</td>
<td>1783-1979</td>
<td>21</td>
<td></td>
<td>43.8</td>
<td>1.41</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>Idkerbergsfältet</td>
<td>1860-1977</td>
<td>11</td>
<td>63.1</td>
<td></td>
<td>0.10</td>
<td>1.57</td>
<td>0.02</td>
</tr>
<tr>
<td>Dannemorafältet c. 1200-1992, 2012-2014</td>
<td>62</td>
<td>32</td>
<td>38.6</td>
<td>1.98</td>
<td>0.00</td>
<td>0.21</td>
<td></td>
</tr>
<tr>
<td>Kölken</td>
<td>not exploited</td>
<td>70</td>
<td></td>
<td>40.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Norrberg-Morberg-Kallmorbergs-fälten</td>
<td>1858-1902</td>
<td>43</td>
<td>19</td>
<td>42.4</td>
<td>0.13</td>
<td>0.05</td>
<td></td>
</tr>
</tbody>
</table>

Table 2. Overview of the most important iron deposits in the Bergslagen district.
Figure 4. Examples of ferrous, mainly iron, ores from Bergslagen (Stephens et al. 2009). a. Apatite-bearing magnetite ore, Grängesberg deposit. b. Actinolite-garnet-magnetite skarn from Mn-poor skarn, Persberg deposit. c. Dannemoreite-knebelite-magnetite in Mn-rich skarns, Dannemora deposit. d. Jasper-hematite layered banded iron formation, Pershytte district. e. Banded iron formation, Striberg deposit. f. Banded dolomitic marble impregnated by hausmannite, Långban deposit. g. Manganite (MnO(OH)), Bölet deposit. Figure f: photo: Erik Jonsson (SGU), sample NRM 31128, copyright: Naturhistoriska riksmuseet. Other photos: Torbjörn Bergman, SGU.
Skarn iron ores

There are more than 3,200 known skarn iron deposits in northern Bergslagen, most of them have been mined in the past. The largest is the Dannemora Fe-Mn- (Zn-Ag-Pb) deposit which consists of about 25 individual ore bodies with both Mn-rich and Mn-poor parts, and some sulphide mineralisations. The deposit is hosted by sedimentary and felsic volcanic rocks, stromatolitic limestone and skarn-altered rocks (Figure 6). The supracrustal rocks, dated at 1894±4 Ma (Stephens et al. 2009) are interpreted to have been deposited in a submarine, saline caldera setting (Lager 2001, Dahlin 2014).

Mining at Dannemora has probably taken place since the 13th century or earlier, although the first written document on the mine dates from 1481 (Rydberg 1981). In the early days, the mining focused on a minor Zn-Pb sulphide occurrence. However, during the 16th century, iron-ore became the main commodity. It is estimated that the total iron ore production until the closure of the mine in 1992 was 54.3 Mt @ 52 % Fe and 2–3 % Mn (Allen et al. 1996a). In 2012, the Dannemora mines was re-opened but the decline in metal prices, especially the iron prices, forced the company into bankruptcy and the mine closed again in 2015.

The Fe deposit at Dannemora consists of massive and stratabound magnetite, often associated with Mn minerals (Figure 3c), diopside, actinolite, chlorite and serpentine (Lager 2001). The ore also contains calcite, dolomite, siderite and...
Rhodochrosite. It was traditionally divided into Mn-rich and Mn-poor varieties, where the Mn content of the skarn was 1–6 % and less than 1 %, respectively. In parts of the deposit, the magnetite itself is manganiferous. Of particular palaeo-environmental interest is the evidence of evaporites in the carbonate sequence (Lager 2001).

Apatite-iron ores

The Fe ores in Bergslagen have, on the basis of their metallurgical properties, been divided into apatite-rich iron deposits, with P contents >0.2 %, and non-apatite iron deposits, with P contents considerably less than 0.2 % (Geijer & Magnusson 1944). Apatite-iron ores in Bergslagen occur in a restricted belt in the northwestern part of Bergslagen (Figure 7). Within this belt there are four major apatite-iron deposits, Grängesberg, Blötberget, Fredmundberg and Lekomberga, each of which contains several ore bodies. The Idkerberget apatite iron ore, a type of ore similar to those in the vicinity of Grängesberg, is located some 20 km to the north, but a proper geological correlation between the Grängesberg subarea and the Idkerberget deposit cannot be made from the available information.

The Grängesberg deposit, the largest apatite-iron ore deposit in Bergslagen, is the third
largest iron deposit in Sweden, only outnum-
bered by the giant Kirunavaara and Malmberget
deposits of northern Sweden. It is also the largest
known metal deposit in the Bergslagen district,
and the southernmost of the known apatite-iron
deposits of Kiruna-type in Sweden. The major ore
body at Grängesberg extends 1500 m along strike
and is up to 100 m wide (Magnusson 1938). It has
been mined down to 650 m depth, but is geophys-
ically constrained to extend down to at least 1500
m. The ore consists of massive magnetite (80 %)
and hematite (20 %) with up to a few percent of
apatite (Figure 4a). Hematite is restricted towards
the structural hanging wall and makes up the
western part of the ore body. The iron ore is host-
ed by sodic-altered dacitic metavolcanic rocks,
dated to 1904 ± 8 Ma (Hallberg et al. 2008). Ore
and host rocks have been intruded by several gen-
erations of mafic dykes and pegmatites (Loostrim1939). Less than 100 m to the west of the Gränges-
berg ore body (Figure 8), similar, but smaller ores
of the Risberg field are found (Hedberg 1907).
They carry somewhat more hematite, whereas the
P contents are similar to those at Grängesberg.
Further to the southwest and along strike from
the Risberg field lies the Ormberg field (Hed-
berg 1907). The ores at Ormberg are dominated
by hematite and the P content is significantly low-
er than at Risberg and Grängesberg. Still further
to the southwest lies the Lomberget field (Hed-
where disseminated apatite-bearing hematite and magnetite ores have been mined in the past. Five kilometres northeast of Grängesberg are the Blötberget and Fredmundberget fields and about 10 km further to the north lies the Lekomberg field (Geijer & Magnusson 1944). The northernmost apatite-iron ore field in the Bergslagen metallogenic district is Idkerberget, 27 km to the north of Grängesberg (Geijer & Magnusson 1944).

Massive sulphide deposits

The Garpenberg Zn-Pb-Ag-Cu deposit is the largest massive sulphide deposit in Sweden with a total tonnage (production, reserves and resources) of 85 Mt @ 5.01 % Zn, 1.83 % Pb, 0.06 % Cu, 51 g/t Ag and 0.1 g/t Au (Boliden 2011a). It consists of several ore bodies that are interpreted as having been emplaced in a dolomitic carbonate rock (Allen et al. 1996a). The mineralizations are
located in the Garpenberg supracrustal inlier which consist of Paleoproterozoic supracrustal rocks bounded by early orogenic granitoids (Figure 9, Jansson & Allen 2011). Multiple folding and faulting have further redistributed and separated them. The oldest part of the Garpenberg mine (Garpenberg Odalfält) was mined for Cu in the 13th century, but Zn later became the most important commodity (Magnusson 1973). In 1972, the Garpenberg Norra mine, about 3 km to the north, was opened. The two mines were eventually connected through an underground drift and may since then be considered as one. During the last decade, several new blind ore bodies have been discovered to the east of the interconnecting drift. These include the Lappberget, Dammsjön, Kasperbo and Kvarnberget discoveries that have increased the reserves and resources significantly (Allen et al. 2010).

The Garpenberg deposit consists of lenses and pods of sulphides with varying proportions of pyrite, sphalerite, galena, chalcopyrite and pyrrhotite hosted in calc-silicate rocks (tremolite skarn) and mica schists (Christofferson et al. 1986). A more Cu-rich mineralisation forms a network of chalcopyrite-pyrite-pyrrhotite-bearing quartz and quartz-fluorite veins in the quartz-mica altered stratigraphic footwall. The deposit is interpreted to be a synvolcanic sub-surface replacement mineralisation in limestone with the Cu mineralisation as a stringer to the Zn-Pb ore (Allen et al. 1996a). In the Garpenberg Norra deposit,stromatolitic structures indicate an organogenic origin for the limestone (Allen et al. 1996a). The Garpenberg supracrustal inlier (Figure 9) also hosts other Zn-Pb deposits and skarn iron deposits in carbonate rocks. At one of these, the Ryllshyttan Zn-Pb-Ag-magnetite deposit, it has been shown that both sulphide and magnetite mineralisation took place by replacement of carbonate rocks (Jansson & Allen 2015). Dating of critical lithologies has shown that the mineralisation is broadly contemporaneous with volcanism in the area, i.e. the mineralisation is epigenetic but broadly synvolcanic.

The Falun Cu-Pb-Zn, Au-Bi-Se deposit is one of the oldest mines in the country and has been in operation for at least a millennium. It has gained an almost mythical importance in both the Swedish society and mining history. Historically, it is without question the most important mineral deposit in Sweden. While it is impossible to obtain any exact figures on total ore production and grades of the deposit, it is estimated that during its lifetime 28 Mt of ore at 2–4 % Cu, 4 % Zn, 1.5 % Pb, 13–24 g/t Ag and 2–4 g/t Au has been mined (Figure 10, Table 1, see Allen et al. 1996a). In the mine, three types
Figure 10. Documented copper production (1612–1894, red) and ore production (1874–1992, green) from the Falu mine. Note the peak in copper production in the 17th century. Data on copper production are from Tegengren (1924), and data on ore production from the “Official Statistics of Sweden, Metal and Mining Industries”.

Figure 11. Geological map of the Falu region, simplified from Stephens et al. (2009). Base metal and ferrous metal data from the Fennoscandian Ore Deposit Database, Eilu et al. (2010).
of mineralisation have been worked: a massive pyrite-dominated sulphide ore with sphalerite, galena and chalcopyrite (Figure 5b); a Cu-Au stringer ore in intensely altered rocks; and younger Au-mineralised quartz veins (Figure 5f). The deposit is hosted by metavolcanic rocks (Figure 11), largely altered to mica schists and so-called ore quartzites, along with marble and skarn (Lasskogen 2010). The rocks were intruded by quartz-phyric subvolcanic intrusions and mafic dykes, then metamorphosed to upper amphibolite facies and experienced several phases of folding and faulting. The deposit has been interpreted as a stratabound volcanic-associated skarn sulphide deposit (Allen et al. 1996a).

An SGU mapping project and a research project at Luleå Technical University are presently ongoing at and around the Falu mine.

About 40 km west of Falun are the Skytt and Näverberg mines. According to available descriptions, they are similar to the Falu deposit, but significantly smaller with a total production amounting to less than 0.1 Mt.

Tungsten deposits

The Bergslagen district has been an important source of tungsten to Swedish industry. More than 100 W-Mo deposits are known from the area and four mines have been in production; the Yxsjöberg, Sandudden, Wigström and Elgfall deposits (Ohlsson 1979). The most important mine was the Yxsjöberg deposit with an estimated tonnage of 5 Mt @ 0.3-0.4 % W. It was mined until 1989 and was then the largest tungsten deposit in Scandinavia (Ohlsson 1980).

All known W occurrences in the area are skarn deposits hosted by Svecofennian felsic metavolcanic rocks and crystalline carbonate rocks (Figure 5d, e). They are mostly concordant or subparallel to bedding in the country rocks. Crystalline carbonate rock is generally found as remnants in the skarn, but is, in places, completely replaced. Scheelite, generally with a significant component of powellite (CaMoO₄), is the only economically important tungsten mineral (Ohlsson 1979). U-Pb dating of titanite that formed during the skarnification of the limestones yields an age of 1789 ±2 Ma, an age that agrees with the age of some of the post-kinematic granitoids from the area (Romer & Öhländer 1994, 1995).

Molybdenum deposits

Molybdenum deposits of “Climax-type”, in which molybdenite is hosted by late-Svecofennian, 1.8 Ga granites and pegmatites, also occur in the area (Figure 5c). Examples are the Bispbergsklack, Pingstabergr and Uddgruvan deposits. All are minor (<0.1 Mt of ore) and have never had any significant economic importance (Hübner 1971).

REE-deposits

Some of the skarn deposits in Bergslagen, in particular those in the Riddarhyttan district, carry high contents of rare earth elements (REE) and have been mined in the past (Geijer 1961, Gustafsson 1990, Andersson 2004). During the period 1860–1919, about 160 t of cerite ore was produced from the Nya Bastnäs mine. In 1923, an additional 825 t was extracted from the waste dumps. The grade of the ore is unknown.

The REE occurrences formed mainly in skarns replacing dolomitic carbonate rocks and are associated with tremolite and talc (Geijer & Magnusson 1944, Geijer 1936, 1961). The main REE-bearing minerals are allanite and cerite, other REE phases include ferriallanite, törnebohmite and bastnäsite (Andersson 2004). The Bastnäs mine is the type locality for the latter. Sulphide minerals, mostly chalcopyrite, bismuthinite and molybdenite are commonly associated with the REE mineralisations. Most authors agree that the REE occurrences in the area are epigenetic replacement deposits. Observations indicate that REE-mineralisation is broadly contemporary with volcanic activity in the region and formed by fluids of igneous origin (Andersson et al. 2013, Jonsson & Högdal 2013).
Lower Palaeozoic U-rich shale, bitumenous black shale (alumshale), conglomerates and phosphoritic shales once covered large parts of the Fennoscandian Shield. Today, remnants of them are found in outliers within the southern and central parts of Sweden, along the eastern margins of the Caledonides, within nappes in the Caledonides, and in the Gulf of Bothnia (Gee & Snäll 1981, Gee et al. 1982a, 1982b, Andersson et al. 1985 and Figure 12). The shales and conglomerates were deposited in basins along the western margin of the Fennoscandian platform (Gee et al. 2008). These basins started to form during the initial breakup of Rodinia in the late Tonian to Cryogenian (c. 850 Ma) (Nystuen et al. 2008). The breakup led to the separation of the Fennoscandian and Laurentian Shields and the formation of the Iapetus Ocean. In the Lower Ordovician, the two continents started to converge, and in mid-Silurian to Lower Devonian times they collided. During the collision, rocks from the Proterozoic basement, the Neoproterozoic to Palaeozoic sediments, including the Palaeozoic shales, and outboard oceanic volcanic terranes were thrust onto each other and the Fennoscandian Shield, forming the present Caledonian nappes (Gee et al. 2008).

Within the Caledonian shale area (Figure 12), two types of sediment-hosted U deposits occur; phosphate (phosphorite) deposits and black shale deposits, as classified according to the International Atomic Energy Association, IAEA. Cuney (2009), who based his classification of U deposits on their genesis, categorize both the black shale and the phosphate deposits as “synsedimentary uranium mineralisations”. These U mineralisations were formed by sedimentation in shallow bays of inland seas that once covered large land areas at times of high sea level. In these inland seas, U was precipitated in phosphorous-rich sediments (phosphorites) or in organic-rich black shale. In both cases organic activity was essential to produce reducing conditions at the bottoms to reduce the solubility of U.

The black shale deposit at Myrviken (Figure 13), with U-V-Mo-Ni-Zn mineralisation, formed
Figure 13. Neoproterozoic to Palaeozoic sedimentary rocks at the eastern rim of and as nappes within the Caledonides. Geological map from SGU’s digital Bedrock Map Sweden, 1:1M, mineral deposit from Eilu et al. (2013).

Table 3. Metal resources in the deposits within the Palaeozoic polymetallic shales (alum shales and phosphorites).

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Reporting</th>
<th>Tonnage (Mt)</th>
<th>Cu (%)</th>
<th>Zn (%)</th>
<th>Ni (%)</th>
<th>Mo (%)</th>
<th>U (%)</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
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<td>Tåsjö</td>
<td>historic</td>
<td>200</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.020</td>
<td>1</td>
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<tr>
<td>Myrviken</td>
<td>NI43-101</td>
<td>3062</td>
<td>0.012</td>
<td>0.042</td>
<td>0.034</td>
<td>0.015</td>
<td></td>
<td>2</td>
</tr>
<tr>
<td>Häggån</td>
<td>not given</td>
<td>1790</td>
<td>0.045</td>
<td>0.032</td>
<td>0.014</td>
<td>0.014</td>
<td></td>
<td>3</td>
</tr>
</tbody>
</table>

in the Middle Cambrian to Lower Ordovician and shares characteristics with black shale deposits elsewhere in Sweden, for example those at Billingen and Kinnekulle in southern Sweden (Andersson et al. 1985 and Figure 12). The geology of the black shales in the Myrviken area is well known due to extensive exploration (27 drill holes) conducted by the State Mining Property Commission (NSG) in the 1970s and early 1980s (Gustafsson 1979, Gee 1979, Gee et al. 1982b).

The stratigraphy at Myrviken consists of Lower Cambrian sandstone resting on top of a weathered Proterozoic basement. The sandstones grade upwards into conglomerates, commonly phosphorite-bearing, which are in turn overlain by a grey siltstone that grades into the black shales of the middle and upper (Furongian) Cambrian. Some black shales also occur in the Lower Ordovician. Near the village of Myrviken the Upper Cambrian black shale was tectonically thickened from 20–30 m to approximately 180 m by Silurian thrusting and folding. The black shales may be traced across the Caledonides in nappes to the Norwegian border (Gee 1979).

The black shale has, on average, 10.4 % organic carbon. In certain, 20–30 m thick layers, the organic C content may reach 13–14 % and U 200–240 g/t (Gee et al. 1982a). Recent exploration in the area (under the name Viken) has resulted in the resource estimate shown in table 3.

The poorly exposed U-rich rocks in the Tåsjö area are composed of phosphatic and calcareous sandstone to greywackes of Lower Ordovician age resting on top of alum shale (Gee et al. 1978, Gustafsson 1979). The U-bearing phosphate deposit at Tåsjö was discovered in the mid-1950s. The area was further explored in the 1970s by the Geological Survey of Sweden and by the companies Stora Kopparberg and Boliden (Armands 1970). In an area of 500 km², approximately 100 holes were drilled, of which c. 25 crossed the U-rich layer. Historic resource estimates for the entire field gave 40 000 t U and 6 Mt P₂O₅, of which c. 1200 t of uranium and 180 000 t of P₂O₅ were considered to be covered by less than 50 m of overburden (Gustafsson 1979). More recent exploration work (Mawson Resouces Ltd., press release January 05, 2006) have not resulted in any further information on the uranium resources for the Tåsjö area. None of the known deposits in the Palaeozoic shales within the area have been exploited.

The area also includes some Pb-Zn mineralisations in Neoproterozoic to Lower Cambrian sandstones. These are essentially similar to the Laisvall Pb-Zn type, but occur within the Caledonian nappes (Christofferson et al. 1979, Chelle-Michou 2008 and Figure 13).

THE GOLD LINE

The gold potential of central Västerbotten county was recognized in the late 1980s when a regional till geochemistry sampling project revealed a southeast-trending gold anomaly, the so called Gold Line (Lindroos 1989, Lindroos et al. 1992, Bark 2008, Figure 14). Intense exploration in the following years resulted in the discovery of several gold prospects, most of which occur in quartz veins or disseminations along deformation zones, i.e. orogenic gold. Two mines have been in production: Blaiken Zn-Au (closed in 2007) and Svartliden Au, still in production, and ore resource estimated have been published for the Fäboliden and Barsele deposits (Table 4).

The bedrock geology of the Gold Line area (Figure 15) consists of metasedimentary rocks and metabasalts of the Bothnian Supergroup, which was intruded by several phases of granitoids (Kathol & Weihed 2005). The metabasalts were emplaced as sills or submarine lava flows. Pillow lavas, splilites and volcanoclastic breccias are common. Granodiorites intruded at an early stage of the orogeny and were deformed together with the supracrustal rocks. Late- to post-orogenic granites (Revsund-type granites) occur
as large massifs in the region.

The Ersmarksberget gold mineralisation, part of the Blaiken Zn-Au deposit, occurs in north-south striking, discontinuous quartz veins in the contact between a tonalitic intrusion and meta-greywackes. The mineralisation is localised within sulphide-rich, carbonaceous metasedimentary rocks. Gold occurs as electrum in free grains within quartz grain boundaries, intergrown with arsenopyrite, and around the arsenopyrite-quartz grain boundaries (Essuka 2011). The Blaiken mine was in operation for only two years and mainly focused on zinc ore. During these two years the mine produced 785 kton of ore at an unknown grade.
The Svartliden Au deposit comprises epigenetic Au and Ag in hydrothermally altered ductile shear zones that have been metamorphosed to mid-amphibolite facies. Minerals detected in the ore include native silver, native gold, electrum, actinolite, grunerite, diopside, amphibole, pyroxene, löllingite, arsenopyrite, native bismuth and pyrrhotite (Eklund 2007). In 2005, the deposit was put into production. Until 2015, 2.84 Mt of ore at 4.5 g/t gold was mined, yielding 368,612 ounces of gold. Today mining has ceased at Svartliden but the concentration plant continues to process low-grade ore (http://www.dragon-mining.com.au).

The mineralisation at Fäboliden is mainly hosted by arsenopyrite-bearing quartz veins within a roughly N-striking, steeply dipping shear zone cutting amphibolite facies volcano-sedimentary host rocks (Bark 2005, 2008). The narrow belt of supracrustal rocks is surrounded by Revsund granites. The gold is fine-grained (2–40 μm) and closely associated with arsenopyrite-löllingite and stibnite, and occurs in fractures and as inclusions in the arsenopyrite-löllingite grains. Gold also occurs as free grains in the silicate matrix of the host rock. Bark (2008) suggested, from his observations at Fäboliden, that favourable places for future exploration for orogenic gold would be areas associated with N-S trending tectonic zones active at around 1.8 Ga.

Gold mineralisation at Barsele predominantly occurs within a medium-grained, highly frac-
tured granodiorite and associated metavolcanic and metasedimentary rocks. Three broad types of mineralisation is recognised: 1) orogenic or mesothermal intrusive-hosted gold mineralisation, 2) high-grade gold-silver-lead-zinc mineralisation hosted by syn-tectonic quartz-sulphide veins and 3) massive sulphide (VMS) where gold is probably mobilised and enriched by a later epithermal mineralisation phase (Orex 2012, technical report).

SKELLEFTE DISTRICT
(Zn-Cu-Pb-Ag-Au-Te)

The Skellefte District and adjacent areas in northern Sweden (Figure 16) is one of the most prominent gold and base-metal districts in the Fennoscandian Shield with c. 150 known precious and base-metal deposits. Around 30 of these have been in production since 1924, when the first mine, at Boliden, was opened (Boliden’s history, at www.boliden.com, dec 2015). Today (2015), there are six active mines: Kristineberg, Maurliden, Maurliden Östra, Renström, Kankberg and Björkdal. Approximately 30 additional deposits with historical mineral resources, as well as modern resource data, have not been exploited yet. The total production from the district (1924–2009) was 105 Mt @ 2.4 g/t Au, 60 g/t Ag, 0.94 % Cu, 4.6 % Zn and 0.5 % Pb. Reported reserves and resources are 7.45 Mt and 25 Mt, respectively, at somewhat lower grades (Boliden Annual Report 2015, http://www.mandalayresources.com/reserves-and-resources/).

The district is defined as a 140 x 50 km, WNW-trending, Palaeoproterozoic (1.96–1.86 Ga) magmatic region formed in a volcanic arc environment. It has a large number of pyritic massive sulphide deposits hosted by metavolcanic rocks (Allen et al. 1996, Kathol & Weihe 2005, Rickard 1986). In this report, rocks in areas to the east of the magmatic rock-dominated Skellefte district proper, which are dominated by metasedimentary rocks, including marble and belong to the Bothnia basin (Kathol & Weihe 2005) have been included since they are host to important gold deposits. The Jörn granitoid complex north of the Skellefte district proper is hosting porphyry copper mineralisation (Weihe 1992a, 1992b) and is, in this report, included in the Skellefte district area.

Most of the deposits are massive to semimassive, complex pyritic Zn-Cu-Pb-Au-Ag occurrences, generally located in the upper parts of the metavolcanic sequence close to overlying metasedimentary rocks (Allen et al. 1996). Economically important exceptions, for example the Boliden and the Kristineberg deposits, occur at lower stratigraphic levels. Orogenic gold deposits, low-grade porphyry Cu, one Ni and numerous sub-economical Au-As quartz vein deposits has also been reported from the district (Grip & Fri-
More than 50% of all the gold produced in Sweden comes from the Skellefte district and in particular from the eastern part of the area where the massive sulfide deposits tend to be more gold-rich and where orogenic gold deposits are found. The individual gold or gold-rich massive sulfide deposits in the area show different characteristics, have different host rocks, and were most likely formed at different times and by different processes. Collectively they point out the gold potential of the area. A selection of the larger deposits is listed in Table 5 and some of them are further described below.

Mineralised metavolcanic rocks of the metallogenic area (the Skellefte Group) are generally overlain by metasedimentary rocks. In detail, however, the stratigraphy of the supracrustal rocks is complex with large variations across and along the district (Allen et al. 1996). The basement to the supracrustal sequences is not exposed. The supracrustal rocks were intruded by a broadly coeval intrusive suite, the Jörn suite, deformed, metamorphosed and subsequently intruded by a younger suite, the so-called Revsund granites (1.82–1.68 Ga). To the south, the Skellefte district is bordered by metasedimentary rocks belonging to the Bothnian basin. To the north, there is a less well-defined boundary against the Arvidsjaur Group, which predominantly consists of continental volcanic rocks, with minor metasedimentary and intrusive rocks (Kathol & Weihed 2005).

Base-metal and gold mineralisation at Boliden (Figures 17–19) occur as massive arsenopyrite ore, massive pyrite+pyrrhotite ore, as veins and as disseminated mineralisation within intensely altered rocks below the massive sulphide ore, and in brecciated parts of the massive sulphide ore (Bergman Weihed et al. 1996). Ore-related hydrothermally altered rocks form a symmetric pattern around the ore with a central zone of intense alteration (andalusite+sericite+quartz) surrounded by less altered rocks (sericite+chlo-
Figure 17. Geological map of the Boliden area from Hallberg (2001). Dots show location of analysed rock samples from outcrops and drill cores and the color indicate the rock classification based on the lithogeochemistry. Coordinates in the Swedish RT90 grid. Vertical black line show location of the profile in figure 19.

Figure 18. Aerial photo of the Boliden deposit with the open pit, head frame and dressing plant. In the upper part of the image the waste rock from the Björkdal gold deposit is seen. Copyright: Boliden Mineral AB.
rite) that in turn grade into a chlorite schist more distal to the ore (Nilsson 1968, Hallberg 2001). The massive ore, the altered rocks and the veins all crosscut the local stratigraphy (Figure 19). The deposit is exceptional among the Skellefte district massive sulphide deposits in terms of its extremely high Au grade (15.5 ppm), high As concentration and intense alteration. These features may suggest an alternative genesis for the deposit, and an epithermal high-sulphidation model, possibly overprinted by an orogenic gold-type event, has been suggested (Bergman Weihed et al. 1996). Recent dating of volcanic host rocks yielded ages of 1894±2 and 1891± Ma (Mercier-Langevin et al. 2013).

The Kristineberg massive sulfide deposit (Figure 20) was discovered in 1918, making it one of the first known deposits in the Skellefte district (Du Rietz 1953). Production started in 1935 and is still on-going. According to official statistics, the mine has, up to 2009, produced 29.3 Mt @ 1.0 % Cu, 3.64 % Zn, 0.24 % Pb, 1.24 g/t Au and 36 g/t Ag. The ore consists of two main massive sulphide zones, the A and the B ores, in addition to the Einarsson zone of Cu- and Au-rich stockwork ores with sulphide lenses in altered and deformed rocks (Årebäck et al. 2005). The ore zones are hosted by hydrothermally altered felsic to intermediate metavolcanic rocks. The deposit was formed at a lower stratigraphic level within the Skellefte Group compared to other deposits in area, including the nearby Rävliden and Rävlidmyran deposits. The mineralisation of the A and B zones is interpreted as synvolcanic massive sulphide type, whereas the Einarsson zone, within strongly altered, andalusite-bearing rocks and rich Au-Cu dissemination, lacks clear evidence of a synvolcanic origin (Årebäck et al. 2005).

Mineralisation in the Adak area was discovered in 1930 during regional exploration conduct-
ed by the Geological Survey of Sweden (Ljung 1974). Mining started in 1934 and the last mine in the area closed in 1976. The Adak field consists of the Adak, Lindsköld, Karlsson östra, Karlsson södra and Brännmyran mines. In total, 6.35 Mt @ 0.80 % Cu, 3 % Zn, 0.1 % Pb and some silver and gold were extracted between 1934 and 1976. The Rudtjebäcken deposit, a few kilometres to the east, was in operation from 1947 to 1975 and produced 2.96 Mt @ 0.92 % Cu, 2.96 % Zn, 0.1 % Pb as well as some Au and Ag (Table 5). Mineralisation in the Adak field consists of chalcopyrite, sphalerite, arsenopyrite, galena, pyrite and pyrrhotite, occurring as massive bodies, veins and disseminations within altered mafic to felsic metavolcanic rocks with carbonate interlayers (Ljung 1974). In addition to base metals and associated low-grade precious metals, the deposits of the Adak field contain Co-bearing pyrite and arsenopyrite. The deposits are also among the most Se-rich in the Skellefte district, and grades of In have been demonstrated (Ljung 1974).

The Åkerberg gold deposit consists of a zone of east–west-trending, subvertical cm-wide gold-bearing quartz veins in a layered gabbro (Mattson 1991, Dahlenborg 2007). The mineralised zone is 10–30 m wide and 350 m long, and can be followed to a depth of c. 150 m. Outside of the mineralised zone, the quartz veins pinch out to altered veinlets and eventually disappear further to the east. The western end of the mineralised zone is not exposed.

The Björkdal gold deposit is located at the contact between an intrusive granodiorite to tonalite and supracrustal rocks belonging to the Skellefte Group. Gold mineralisation consists of cm-to-m-wide, subvertical, auriferous quartz veins within the granodiorite (Weihed et al. 2003, Gold Ore Resources 2011). The veins mainly trend NE and NNE. The gold occurs as both free milling and associated with pyrite.

The Åkulla Östra deposit lies in an area with several other small massive sulfide deposits that have been mined in the past. It was known at an early stage that the deeper parts of the Åkulla Östra mine carried auriferous quartz veins in altered rocks (Grip & Frietsch 1973). Further drilling in the area revealed an Au-Te mineralisation below the massive sulfide deposit. Underground mining of the Au-Te mineralisation commenced with a relatively low base-metal grade that are found in the central Skellefte district.
Figure 21. Drillcore box from Maurliden ddh 58103, 46.27 – 62.57 m showing massive sulfide mineralisation with some shale between 51.64-52.99 m. The average composition of the drillcore box is 0.10 % Cu, 5.1 % Zn, 0.4 % Pb, 0.99 g/t Au and 46.7 g/t Ag (Drillcore protocol Maurliden bh: 58103). 

a. Natural colour (RGB) photo

b. False colour composite image in long wave infrared light. Three infrared bands (R=8611 nm, G=10022 nm and B=11810 nm) compose the image and they represent a color rendition of data from sensors measuring across the long wavelength portion of the infrared spectrum. The image has not been further processed to a mineral map but the orange colored core probably represents the shale intercalation between 51.64-52.99 m. The following section, 53.01-54.17, dark grey in the image is a very zinc-rich part of the core with 21 % Zn.

Color images and false color composite images of the full core, other cores from Maurliden as well as other drillcores from Swedish mineralisation can be seen at www.sgu, choose MapViewer; Drill cores. The location of the drill core in relation to the ore is shown in Montelius et al. 2004.
in 2012 through a tunnel from the then closed Kankberg mine. The new mine is named Kankberg nya in this publication in order to avoid confusion with the old Kankberg mine.

The Tallberg deposit, located to the north of the Skellefte District proper, is interpreted as a Palaeoproterozoic porphyry copper deposit hosted by a 1.9 Ga granitoid (Weihed 1992a, 1992b). The deposit is of low grade, 0.27 % Cu and 0.2 g/t Au, but the large tonnage could make it mineable in the future.

The Holmtjärn gamla (Old Holmtjärn) deposit was found in the early 1920s and was mined out in two years. The upper parts of the ore were strongly weathered and carried gold grades of more than 1000 g/t. The massive arsenopyrite and pyrite ore beneath the weathered zone also showed high gold grades (Grip & Frietsch 1973). Although no detailed description exists, it seems from the available information that Holmtjärn gamla in many ways resembles the Boliden deposit.

Throughout the Skellefte district, several gold- and arsenopyrite-bearing quartz veins have been found (Högbon 1937, Grip & Frietsch 1973). In a few places, they have been mined on a small scale in the past, but small tonnage and scattered gold make them uneconomic.
Sandstone-hosted Pb-Zn deposits of the Laisvall-type are found in Neoproterozoic to Lower Cambrian sandstones exposed along the eastern rim of the Caledonian orogen and in nappes of the same rocks that have been thrust eastward during the Caledonian orogeny (Figure 22). The majority of the deposits are found in Sweden (Grip & Frietsch 1973, Rickard et al. 1979, Casanova 2010, Saintilan et al. 2014, 2015), but some occur in Norway (Björlykke and Sangster, 1981). Mineralisation consists of galena and sphalerite filling up pore space in the sandstones forming stratabound mineralisations. A characteristic feature of the Laisvall-type deposits in Sweden and Norway is the strongly radiogenic composition of galena (Rickard et al. 1981).

The largest and best-described of the sandstone-hosted Pb-Zn deposits in Scandinavia is the Laisvall deposit (Rickard et al. 1979, Wilden 2004, Casanova 2010, Saintilan et al. 2015 and references therein, Figure 23). The deposit was discovered in 1939, the development of the mine started in 1941 and the first ore was produced in 1943 (Willden 2004, Berghshantering 1943). When the mine closed in 2001 it had produced 64,256 Mt of ore with an average grade of 4.0 % Pb, 0.6 % Zn and 9 g/t Ag (Willden 2004).

Pb-Zn mineralisation at Laisvall is hosted in two distinct, nearly horizontal sandstone sequences, the Lower Sandstone-ore horizon (Kautsky Ore Member) and the Upper Sandstone-ore horizon (Nadok Ore Member), Figure 24 and 25. These two ore-horizons are separated by the barren Middle Sandstone (Tjalek Member). In addition, there is minor mineralisation in sandstones of the Neoproterozoic Ackerselet Formation stratigraphically below the main mineralisation (Casanova 2010). The Pb-Zn mineralisation at Laisvall has been dated to Middle Ordovician (467 ± 5 Ma, Saintilan et al., 2015), indicating that it took place about 100 million years after the deposition of the host rocks.

The autochthonous sedimentary sequence of the Laisvall Group, which also includes the Grammajukku Formation on top of the mineralised sandstones, rests unconformably on a weathered surface of Palaeoproterozoic (1.87–1.66 Ga) granitoids to syenites. The Neoproterozoic and younger sedimentary successions hosting the Pb-Zn occurrences of Laisvall type were formed during crustal extension and formation of sedi-
Figure 23. Geological map of the Laisvall area, from Lilljequist 1973. Deposit information from Eilu et al. 2013 and SGUs mineral resources database.
mentary basins along the margins of continental Baltica (Gee et al. 2008).

Saintilan et al. (2015) showed that the basement structures underneath the mineralised sandstones played an essential role, both in localizing the sediments forming the host rocks and as feeders for Pb-Zn fluids. Sulphur isotopes indicate that most of the sulphur comes from thermogenic sulphate reduction of seawater sulphur but also a contribution from pyrite in the overlying black shales and reduced sulphur brought up with the Zn-Pb fluids (Saintilan et al. 2014). Nobel gas isotopes in fluid inclusions in ore minerals indicate low temperature crustal fluids, the halogen compositions of fluid inclusions suggest that the fluids was derived from evaporation of seawater beyond halite saturation (Kendrick et al. 2005).

Two similar, unexploited Pb-Zn deposits have been identified in the vicinity of Laisvall. They are Maiva, with about 1 Mt of ore, 7 km to the northeast of Laisvall, and Niepsurt with 1.75 Mt of ore located less than 10 km to the south (Lilljequist 1973, Figure 23 and Table 6). In addition, there are another 15 known sandstone-hosted Pb-Zn deposits in the Laisvall area (Grip & Frietsch 1973).

<table>
<thead>
<tr>
<th>Deposit</th>
<th>When mined</th>
<th>Resources (Mt)</th>
<th>Mined (Mt)</th>
<th>Ag (g/t)</th>
<th>Pb (%)</th>
<th>Zn (%)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Laisvall</td>
<td>1943–2001</td>
<td>1.00</td>
<td>64.3</td>
<td>9</td>
<td>4.0</td>
<td>0.6</td>
<td>1</td>
</tr>
<tr>
<td>Maiva</td>
<td></td>
<td>1.75</td>
<td>8</td>
<td>3.2</td>
<td>0.3</td>
<td></td>
<td>3</td>
</tr>
<tr>
<td>Niepsurt</td>
<td></td>
<td>1.75</td>
<td>8</td>
<td>3.2</td>
<td>0.3</td>
<td></td>
<td>3</td>
</tr>
</tbody>
</table>

Table 6. Sandstone-hosted Pb-(Zn) deposits in the Laisvall area.

Figure 25 a. Natural colour (RGB) photo of drillcore box from Laisvall ddh 34, 92.97-110.3 m. The upper part of the core box contains core from the mineralized Upper Sandstone-ore horizon (Nadok Ore Member), the uncut core in the central parts of the core box contains core of the barren Middle Sandstone (Tjalek Member) and the lower part of the core box contains core of the mineralized Lower Sandstone-ore horizon (Kautsky Ore Member).

b. Magnification of the lower right corner of the core box shows the typical speckled appearance of Pb-Zn mineralisation of Laisvall type. Colour photos and false colour composite image of infrared imaging of the full core, other cores from Laisvall as well as other drillcores from Swedish mineralisation can be seen at www.sgu, choose MapViewer; Drill cores.
Several quartz-banded iron deposits (Proterozoic banded iron formation, Frietsch 1997, Grip & Frietsch 1973) occur in the Kallak area (Figure 26). The host rocks to the deposits consist of felsic to intermediate volcanic rocks and metasedimentary rocks (banded gneisses) belonging to the Kiruna-Arvidsjaur Group (Porphyry group) with an age of 1.88-1.86 Ga (Bergman et al. 2001). The supracrustals have been intruded by several generations of both mafic and felsic compositions.

At the Kallak (a.k.a. Björkholmen) deposit, the largest and best-known deposit in the area, magnetite and hematite occur interlayered with quartz and feldspar with accessory hornblende, diopside and chlorite in a 1 km long and up to 300 m wide zone (Eriksson 1983). Layers of skarn, with garnet and epidote, locally form thick layers between the iron mineralisation and the host rocks (Frietsch 1997).

The iron-ore potential of the area was recognized in the 1940s by SGU when the Kallak deposit was discovered and drilled. Continued exploration in the 1970s found several deposits similar to Kallak. These include, from south to north, Akkihaure, Tjårovarets, Southern Parkijaure, Parkijaure, Southern Kallak, (Kallak), Maivesvare, Åkosjägge and Pakko. Historic resource estimates based on ground magnetic and gravimetric surveys exist for four of the deposits, modern resource estimates according the JORC code exist for Kallak and Southern Kallak (Table 7). Documentation from exploration work in the Kallak area includes several drillcores and ground magnetic surveys (digitised), mainly from SGU exploration, but also from exploration companies.

Exploration in the late 1970 revealed copper mineralisation hosted by intrusive rocks at Iekelvare to the south-west of the Kallak area (Weihed 2001, Sundberg et al. 1980). Several objects in the area were explored and drilled around 2005, but with poor results.

None of the deposits in the Kallak area have been exploited so far, but the company Beowulf plc. is currently developing the Kallak deposit (www.beowulfmining.com)

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Resources (Mt)</th>
<th>Standard</th>
<th>Fe (%)</th>
<th>Reference to deposit size</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kallak</td>
<td>144.1</td>
<td>JORC</td>
<td>35</td>
<td>Press release 3 April 2013 Beowulf Mining</td>
</tr>
<tr>
<td>Åkosjägge</td>
<td>10</td>
<td>JORC</td>
<td>23</td>
<td>Hannans Reward annual report 2012</td>
</tr>
<tr>
<td>Akkihaure</td>
<td>12</td>
<td>historic</td>
<td>25</td>
<td>Frietsch 1997</td>
</tr>
<tr>
<td>Parkijaure S</td>
<td>6</td>
<td>historic</td>
<td>35</td>
<td>Frietsch 1997</td>
</tr>
<tr>
<td>Parkijaure</td>
<td>23</td>
<td>historic</td>
<td>38.5</td>
<td>Frietsch 1997</td>
</tr>
</tbody>
</table>

Table 7. Resources and Fe-content of the iron deposits in the Kallak area with references.
Figure 26. Geological map of the Kallak area simplified from SGUs digital map Bedrock of Sweden 1:1M. The mafic and felsic metavolcanic rocks of the Kiruna-Arvidsjaur Group (Porphyry group) and hosting the iron deposits are framed in the legend. The iron deposits, from Eilu et al. (2013), are marked with blue dots where the size of the dot indicate the size of the deposit, see table 7.
The Aitik Cu-Au-Ag mine is located some 15 km southeast of Gällivare and is presently Europe’s largest open-pit copper producer (Figs. 27 and 28). From a modest production of less than 2 Mt during the first year of operation (1968) to the present (2013) production of c. 39 Mt (New Boliden 2014) the deposit has produced 632 Mt @ 0.35 % Cu, 0.18 g/t Au and 3.4 g/t Ag (New Boliden 2014). Reported reserves are 756 Mt @ 0.22 % Cu, 0.15 g/t Au, 1.5 g/t Ag and 24 g/t Mo, while measured and indicated resources are 1643 Mt @ 0.16 % Cu, 0.10 g/t Au, 1.0 g/t Ag and 23 g/t Mo (New Boliden 2014).

The ore at Aitik is hosted by a biotite-sericite schist or gneiss and amphibole-biotite gneiss (Zweifel 1976, Monro 1988). The precursor to the altered, metamorphosed and deformed host rock is a ca. 1.9 Ga volcanosedimentary sequence formed during the Svecokarelian orogeny (e.g., Witschard 1996, Wanhairen et al. 2006). Stratigraphically, the sequence belongs to the Muorjevaara Group (Martinsson and Wanhairen 2004) and is correlative with the regional Porphyrite Group of Bergman et al. (2001). A summary of the local stratigraphy is presented in Lynch et al. (2015).

The Mineralisation consists of veinlet and disseminated chalcopyrite, pyrite, pyrrhotite, molybdenite and magnetite. Accessory minerals include bornite, chalcocite, malachite, sphalerite, galena, arsenopyrite, scheelite, uraninite andapatite. Native gold, electrum and amalgam mainly occur in fractures, along grain boundaries and as inclusions in sulphide and silicate
minerals (Sammelin (Kontturi) et al. 2011). In the footwall, a deformed quartz monzodiorite with sub-economic Cu grades has yielded a U-Pb zircon age of 1887 ± 8 Ma (Wanhainen et al. 2006). Reported Re-Os and U-Pb ages for hydrothermal minerals have identified mineralisation-alteration stages at ca. 1.88, 1.85, 1.80 and 1.78 to 1.73 Ga, and suggest an episodic and protracted metallogenic evolution (Wanhainen et al. 2005). Both the monzodioritic stock and the host volcanosedimentary rocks were affected by ore-related potassic alteration. In general, Aitik is interpreted as a Palaeoproterozoic porphyry Cu deposit formed at ca. 1.89 Ga that was subsequently overprinted by an iron oxide-copper-gold-style mineralization event some 100 million years later (Monro 1988, Wanhainen 2005). A deposit similar to Aitik with respect to the style of mineralisation, grade and host rocks, but of much smaller size (ca. 9.5 Mt), is located at Liikavaara East, about 4 km to the east of Aitik (Sammelin 2011).

Further to the north of Aitik, the NNE-trending Nautanen Deformation Zone (NDZ; Witschard 1996) hosts several relatively small Cu-Au prospects (typically < 3 Mt), including the Nautanen and Liikavaara deposits (Martinsson and Wanhainen 2004, 2013 and Figure 29). The mineralisation occurs as disseminations, veinlets and local semi-massive lenses of chalcopyrite, bornite and pyrite (± magnetite). Deformed and altered host rocks in the area are lithologically and geochemically correlative with the wall rocks at Aitik (i.e., Muorjevaara Group meta-volcanosedimentary rocks; Lynch et al. 2015). Smith et al. (2009) report U-Pb ages ranging from ca. 1.79 to 1.78 Ga for hydrothermal alteration at the Nautanen deposit and inferred temporal and genetic links between deformation, granitic magmatism, fluid mobilisation and Cu-Au mineralisation. To the east of the NDZ, the Ferrum and Friedhem prospects are examples of mainly quartz vein-hosted Cu-Au mineralisation.
The **Malmberget** apatite iron ore deposit (Figure: see chapter Aitik-Nautanen, figure 27) is the second largest iron deposit in Sweden with a combined total production plus reserves and resources of 1177 Mt (Mineral statistics of Sweden and www.lkab.com). The deposit consists of several ore bodies over an area of around 15 square kilometers (Figure 30). In the western and northern part of the Malmberget ore field, the ore forms an almost continuous horizon, whereas the eastern part is made up of several isolated lenses of iron ore. (Figure 30, Bergman et al. 2001). One of the ore bodies, the blind Printzsköld ore body at 700 m depth and below have been investigated in detail by Debra (2010).

The dominant ore mineral in the Malmberget iron ore bodies is magnetite but hematite-rich ore becomes more frequent in the western parts of “Stora Malmlagret” at the ore bodies Välkommnan, Baron, Johannes and Skåne (Bergman et al. 2001).

Host rock to the deposit mainly consists of metamorphosed and deformed volcanic rocks of rhyolitic to dacitic composition belonging to the Kiruna-Arvidsjaur Group (Porphyry Group) and with an age of 1.88-1.86 Ga (Bergman et al. 2001). The felsic metavolcanic host rocks are commonly rich in K-feldspar, whereas albite-rich rocks occur locally as hosts to ore. At some places amygdules have been observed in the host rocks indicating an extrusive character of the rocks. The mafic rocks in the area mainly occur close to the iron ores. They are interpreted to be dykes, sills and possibly also extrusives. To the north the ore-bearing metavolcanic rocks are intruded by a large granite-pegmatite (Lina granite) with an age of 1.85-1.75 Ga. Recrystallisation of the metavolcanic rocks increases towards the granite-pegmatite. Granite dykes and pegmatites are common in the iron ore and in their host rocks (Bergman et al. 2001, Witschard 1996).

The Malmberget deposit and metavolcanic host rocks are highly metamorphosed. Ductile deformation has formed the present shape of the deposit with a large scale fold with a fold axis dipping moderately to the SSW and intense stretching of ore bodies parallel to the fold axis (Geijer 1930, Grip and Frietsch 1973, Bergman et al. 2001). Younger dikes, believed to belong to the younger granite-pegmatite association, are similarly deformed suggesting that deformation took place after ca. 1800 Ma (Bergman et al. 2001).

To the west of the ore bodies making up the Malmberget ore field there are some smaller deposits believed to be of the same type of iron ore as Malmberget. These include the Sikträsk and the Norregruvan deposits, occurring in enclaves of supracrustal rocks in the granite-pegmatite (Lina granite, Grip and Frietsch 1973, Geijer 1930).

The deposits in the Malmberget area were probably known already in the 17th century, but it was not until technical development (the Thomas process for treating P-rich iron ore) and infrastructure, in the form of a railway to a port in Luleå at the Baltic Sea and later on a railway to the Atlantic port in Narvik, Norway, that large-scale iron ore production commenced. From 1885 to 1905 the production of lump ore increased from less than 100 tons/year to 1 million tons/year.
Figure 30. Geological map of the Malmberget area, from Bergman et al. 2001. All rocks belong to the Porphyry group, except the rocks of the Granite-pegmatite association.

Figure 31. Aerial photo of the Malmberget area. Photo Fredric Alm, LKAB
Northern Norrbotten hosts a large number of apatite iron ores of the Kiruna type (Frietsch 1997, Bergman et al. 2001 and Figure 32). The largest, and economically most important deposit of this type is the Kirunavaara iron deposit, also being the “type locality” for apatite e-iron ores, but there are several other large-scale iron ore producers as well as unexploited deposits of this type in the area (Bergman et al. 2001, Frietsch 1997, Grip & Frietsch 1973).

The iron ore potential of the Kiruna area has been known since the 17th century, but large-scale mining did not start until the arrival of infrastructure, in the form of railways to the Baltic Sea and the Atlantic coast during the first years of the 20th century (Frietsch 1997). This, together with new metallurgical methods that made it possible to produce iron and steel from apatite-bearing iron ore (the Thomas process) opened up the area for mining. From modest and intermittent mining in the 19th century through the last 110 years of large-scale mining, 1714 Mt of apatite iron ore @ c. 60 % Fe have been produced from Northern Norrbotten (including the Malmberget deposit, see the Malmberget chapter). Present day reserves and resources of apatite e-iron ores in Norrbotten is 2372 Mt (Data from LKAB 2014 and Official statistics of Sweden).

The apatite-iron deposits in the Northern Norrbotten area occur in, or have a spatial relation to the Porphyry Group (Figure 32), a bimodal sequence of mafic to felsic metavolcanic rocks and metasedimentary rocks, or in a few cases in the underlying Porphyrite Group, which predominantly consists of meta-andesites (Bergman et al. 2001). These rocks were deposited between

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**Figure 32. Distribution of supracrustal rocks belonging to the Porphyry Group (also Arvidsjaur-Skelletfje Group) in Northern Norrbotten, from Bergman et al. 2001. Yellow colour for felsic metavolcanic rock, green colour for mafic metavolcanic rock. Data for deposits from Eilu et al. 2013.**
In the Kiruna area, the Porphyry Group is overlain by the Lower Hauki Group, which consists of quartz-sericite altered metavolcanic and metasedimentary rocks (Parak 1975a, Bergman et al. 2001). In most cases, magnetite is the dominant mineral and with variable amounts of apatite, commonly F-apatite, and actinolite, albite and scapolite as gangue. A few deposits are hematite-dominated and there the gangue consists of apatite, quartz and carbonates. In general, the hematite-dominated deposits are found higher in the stratigraphy of the Porphyry Group. The phosphorous content of the Kirunatype deposits in Sweden is c. 0.5-1.0 % P but smaller sections of ore bodies as well as the host rocks can contain several percent of P. The contents of Ti and S are very low, in contrast to nelsonites, rich in titanium and skarn iron ores which generally are sulfide-bearing. Most of the apatite iron deposits are also enriched in rare earth elements (REE), which are mainly concentrated in the apatite (Frietsch & Perdahl 1995).

The Kirunavaara deposit is a tabular ore body that can be followed for about 5 km along strike, is up to 100 m thick and has been shown to extend to a depth of more than 1300 m (Parak 1975a, Bergman et al. 2001, Figure 33). Calculations based upon geophysical measurements indicate that the ore body continues below a depth of 1500 m. The ore is at the contact between a thick sequence of trachyandesitic lava flows and the overlying rhyodacitic pyroclastic rocks (Figure 33). The entire sequence strikes N–S and dips steeply to the E. The massive magnetite-apatite ore grades into magnetite-actinolite breccias towards the wall rocks. The P content of the ore varies and shows a bimodal distribution of either <0.05 % or >1.0 %. In the faulted southern end of the Kirunavaara deposit, three small ore bodies, Konsuln, Sigrid and Viktor, are situated. To the north, the mineralisation may be followed at depth under Lake Luossajärvi (today partly drained). Two kilometres further north is the Luossavaara deposit, showing the same characteristics as Kirunavaara. At the surface, Luossavaara can be traced for 1200 m along strike. Deposits occurring further to the north, Nukutus and Henry, are so-called Per Geijer ores (Geijer 1919, Frietsch 1997). The Per Geijer ores differ from the larger Kirunavaara and Luossavaara deposits, in size, dominant iron oxide and gangue mineralogy, alteration style and in stratigraphic position. In general, they contain more apatite and hematite. Locally, hematite is the dominant oxide. Gangue consists of quartz and sericite and, in places, also carbonates and albite. These deposits occur in the upper parts of the Porphyry Group, stratigraphically above the Kirunavaara and Luossavaara deposits. The Rektorn and Haukivaara deposits, to the east of Kirunavaara, are also Per Geijer ores. The blind Lappmalmen deposit, discovered in the 1960s, shows the same characteristics as the Per Geijer ores (Parak 1969).
From modest and intermittent mining at Kirunavaara in the 19th century through the last 110 years of large-scale mining, 1034 Mt of iron ore @ c. 60 % Fe have been produced. Present day reserves and resources are 682 Mt and 304 Mt, respectively, giving a total tonnage of the Kirunavaara deposit of more than 2 Gt (LKAB 2014, Official statistics of Sweden, Metal and Mining Industries). In the area there are several other historical producers (Figure 33, Table 8), including Luossavaara which produced 21.2 Mt between 1858 and 1985, Nukutus (5.3 Mt) and Haukivaara (2.4 Mt). A few km to the E of Kirunavaara lie the Tuolluvaara deposit, in operation between 1902 and 1982 with a total production of 25.4 Mt. The Mertainen deposit is located 30 km to the ESE. During the short lifetime of this mine, from 1956 to 1959, the deposit produced 0.4 Mt of iron ore but the mine is planned to be re-opened soon. A large unexploited resource, the blind Lappmalmen deposit, is located 2 km east of the Luossavaara deposit (Figure 33).

Leveäniemi, 40 km southeast of Kirunavaara is the third largest apatite-iron deposit in Northern Norrbotten with a total tonnage (reserves and mined tonnage) of 168 Mt @ 55.4 % Fe and 0.45 % P (Frietsch 1966, 1997, LKAB 2010). The Leveäniemi deposit was mined from 1964 to 1982 but was recently (2014) re-opened. The deposit is dominated by massive magnetite ore, massive hematite ore and calcite-rich magnetite ore. Large volumes of magnetite breccia (not included in the resources) occur in an up to 100-m-wide zone in the surrounding biotite schist. The contacts between massive ores and ore breccias are mostly distinct. The nearby Gruvberget apatite-iron deposit (Frietsch 1966), geologically similar to Leväniemi, was recently (2010) put into production by LKAB. In older days, vein-hosted copper was mined at Gruvberget. During intermittent production between 1644 and 1785, about 1740 tons of raw copper was produced (Tegengren 1924).

The Kiruna area also hosts small epigenetic Cu deposits. These are mainly hosted by the Greenstone Group rocks, which form the basement to the Porphyrite and Porphyry groups. The only deposit of this kind that has been mined in recent times is Pahtohavare (Martinsson 1997), which is immediately outside the Kiruna metallogenic area. The Rakkurijärvi Cu deposit, a few km S of Kirunavaara, has recently been subject of detailed exploration (Smith et al. 2007). Other deposits of this kind include Tjärrojäkka, where part of the Cu occurrence is hosted by an apatite-iron ore (Edfelt 2007), and the Sierkavara (Pikkujärvi) deposit (Weihed 2001, Hedin 1988).

The genesis of Kiruna-type deposits has been
discussed for almost a hundred years without any conclusive evidence for any of the ore-forming models. The discussion on the genesis for this type of deposits started in Southern Sweden through studies of the Grängesberg apatite e-iron ore in the Bergslagen district. In the first modern description of the Grängesberg deposit Johansson (1910) included a model for the genesis of the ores. He suggested that the iron deposits together with the host-rocks to the deposit were formed by in situ differentiation, that is the iron ore was segregated from the surrounding metavolcanic rocks. Looström (1929a, 1929b, 1939) argued against that interpretation and favoured an intrusive origin for the deposits. He also suggested that the iron deposits together with the host-rocks to the deposit were formed by in situ differentiation, that is the iron ore was segregated from the surrounding metavolcanic rocks. Looström (1939) was the first to point out the similarities between Grängesberg and the large apatite-iron deposits in northernmost Sweden and his statements was based on the model for the large apatite-iron deposits at Kiruna, Gällivare, and Tuollavara in northern Sweden (Geijer 1910, 1918, 1920, 1930.)

From the early 1970s and onwards the focus for the discussions on the genesis of apatite-iron ores shifted towards the deposits in Northernmost Sweden and to similar deposits around the world. Parak (1975a, 1975b) argued against the magmatic model for the Kiruna deposit and suggested a sedimentary origin. This was criticised by Frietsch (1978, 1984) and others (Gilmour 1985, Wright 1986). In 1994 the discussion continued by a publication by Nyström & Henriquez (1994) in which they showed the similarities between the Kiruna deposit and much younger apatite-iron deposits in Chile. The discussion continued by contributions by Nyström & Henriquez (1995), and Bookstrom (1995). Throughout these years there were also publications on other apatite-bearing iron deposits in US and in Canada (Panno & Hood 1983, Hildebrand 1986), both in favour of a hydrothermal genesis for the apatite-iron deposits they described.

In the early 1990s Hitzman et al (1992) published a paper in which they discussed the genetic relations between apatite-iron deposits of the "Kiruna-type", the giant Olympic Dam Cu-U-Au-Ag deposit, and other deposits. They proposed that these deposits, characterised by their age, tectonic setting, mineralogy, and alteration should be referred to as Iron-Oxide Cu-U-Au-REE deposits. Later on the name became Iron Oxide Copper Gold (IOCG) deposits. In their model, in which apatite-iron deposits of the "Kiruna-type" make up a subset, the mineralisation is caused by hydrothermal processes. This new concept has triggered a lot of exploration for IOCG-deposits around the world and in Sweden.

To the genetic discussion on apatite-iron deposits Sillitoe & Burrows (2002) contributed with a description of the El Laco apatite-iron deposits in Chile. There they argued for a hydrothermal replacement genesis for the deposit, a model that was questioned by Henriquez et al (2003).

<table>
<thead>
<tr>
<th>Mine</th>
<th>In operation</th>
<th>Ore production (Mt)</th>
<th>Resources (Mt)</th>
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<td>1033.8</td>
<td>986</td>
</tr>
<tr>
<td>Malmberget *</td>
<td>1845-</td>
<td>561.6</td>
<td>580</td>
</tr>
<tr>
<td>Leveäniemi</td>
<td>1964-1982, 2014-</td>
<td>57.3</td>
<td>332</td>
</tr>
<tr>
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<td></td>
</tr>
<tr>
<td>Luossavaara</td>
<td>1846-1985</td>
<td>21.2</td>
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</tr>
<tr>
<td>Gruvberget</td>
<td>1860-1892, 2010-</td>
<td>6.4</td>
<td>81</td>
</tr>
<tr>
<td>Nukuts</td>
<td>1964-1986</td>
<td>5.3</td>
<td></td>
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<td>Haukivaara</td>
<td>1965-1972</td>
<td>2.4</td>
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</tr>
<tr>
<td>Mertainen</td>
<td>1956-1959</td>
<td>0.4</td>
<td>393</td>
</tr>
</tbody>
</table>

Most of the ore contained more than 60 % Fe and c. 0.58 % P. Data from LKAB (2014) and Official statistics of Sweden.

* The Malmberget deposit is discussed elsewhere in this publication.
In Paleoproterozoic time, at around 2.44 Ga, a rifting event commenced in Northern Fennoscandia with intrusion of ultramafic to mafic rocks. This was followed by deposition of sediments, mafic volcanics and carbonate rocks in a rift-related tectonic setting (Figure 35). Numerous intrusions of mafic dykes and sills of the same age are also found in the area. The deposition of these rocks, the Karelian supracrustal rocks, occurred approximately at 2.4 to 1.96 Ga ago. The Karelian supracrustal rocks are found in Northern Sweden, Northern Norway, Northern Finland and in Northwestern Russia (Bergman et al 2001).

In Northern Sweden the Karelian supracrustal rocks are subdivided into the older Kovo Group (2.4-2.3 Ga) and the younger Greenstone Group (2.3-1.96 Ga). In the Kiruna area the Kovo Group consist of conglomerates, quartzites and other sediments unconformably overlying the Archaean basement (Figure 36). The stratigraphy of the Greenstone Group has been described from many places in Northern Sweden as well from Finland and Norway and there are several names for this group, i.e. Kiruna greenstones, Kiruna greenstone group, Vittangi greenstone group, Veikkavaara greenstone group, Käymäjärv group, Kolari greenstones, Greenstone formation, Iron ore formation, Schist formation etc. (Bergman et al. 2001) In the following text we use the term Greenstone Group to describe these rocks.
The main rock types in the Greenstone Group are metabasalt, graphite bearing meta-argillites, crystalline carbonate rocks and ultramafic rocks (Martinsson 1997, Bergman et al. 2001, Grigull & Jönberger 2014).

The Greenstone Group hosts several types of mineralisations (Frietsch 1997, Eilu et al. 2013, Figure 35, Table 9, Table 10). Most common are stratabound skarn iron formations, copper mineralisations, copper-cobalt mineralisations, graphite deposits and carbonate deposits.

For the convenience of the reader the deposits in the Norrbotten Greenstones are described from west to east, divided into three areas; western, central and eastern area. This division does not refer to any major difference in composition or stratigraphy in the Greenstone Group going from west to east.

**Western area**

The most important deposit in the western part of the Greenstone Group is the Viscaria Cu deposit. It consists of several stacked units showing a variation from magnetite-bearing Cu-rich sulphide ore in the A zone to sulphide poor magnetite ore in the B and D zones (Figure 37; Martinsson 1997). The economically important A zone occurs in a marble between two units of black schist, on top of a volcaniclastic unit and immediately below pillow lavas. It is capped by an extensive thin chert unit. The ore consists of fine-grained chalcopyrite, magnetite, pyrrhotite and minor sphalerite. The ore minerals are disseminated or form thin intercalations and semi-massive accumulations. According to Martinsson (1997), the ore was formed by fissure-controlled exhalative events in a fault-controlled basin. The deposit was in operation 1982-1997 (Table 10). Ongoing exploration has resulted in new resource data for the Viscaria deposit including both copper and iron resources (http://avalonminerals.com.au/).

The Tjavelk deposit is a magnetite skarn iron ore, which is approximately 600 m long and up to 30 m wide. It is estimated to have 6.8 Mt @ 35 % Fe, 3.6 % S and 0.12 % Cu (Frietsch 1997). The skarn is composed of tremolite-actinolite...
and serpentine with a banded pattern in places. Sulfide minerals are pyrrhotite, pyrite, chalcopyrite and some pentlandite. Apatite is relatively common both as impregnations and as schlieren in the magnetite ore. The average phosphorus content is 1.3 % P, the highest found in a skarn iron ore in northern Sweden.

The iron ore zone at Sautusvaara is approximately 2500 m long and strikes NW–SE. The deposit consists of two ore bodies separated by a fault (Hallgren 1970). The smaller northern ore body has been estimated to have 13.3 Mt @ 42.1 % Fe, the southern ore body 42.1 Mt @ 37.2 % Fe (Hallgren 1970). The mineralisation consists of bands of magnetite and skarn minerals (diopside and tremolite). In places, chlorite and biotite are abundant. The dominant sulfide mineral is pyrite, generally occurring as fissure veins. At least two generations of pyrite occur and the content of cobalt in the coarse grained pyrite varies between 1.1–1.6 % Co. Traces of chalcopyrite and pyrrhotite have been detected. The rock underlying the iron ore zone is a well-stratified scapolite-diopside-biotite-bearing sediment. This unit is underlain by graphite-bearing schists and minor marble with skarn. These rocks are stratigraphically situated in the upper part of the Greenstone Group.

The Pahtohavare deposit SW of Kiruna comprises three epigenetic Cu-Au ore bodies (Martinsson 1997). Two of them have been mined (the Southern and the Southeastern ores). The ore bodies are located in an antiformal structure, bordered towards south by a shear zone (Martinsson 1997). Southern Pahtohavare is the largest ore body, with a maximum length of 270 m and a thickness of up to 25 m. It occurs in an altered black schist unit close to a thick mafic sill. Early albitionisation of tuffite along the contact of the sill was later overprinted by ore-related alteration. Adjacent black schist was albitated and its graphite replaced resulting in a rock commonly called ‘albite felsite’. The main ore minerals, chalcopyrite and pyrite, mainly form veinlets and breccia fill in the albite rock. The third ore body at Pahtohavare, Central, shows indications of supergene control of mineralisation and consist mainly of secondary Cu minerals.

Central area
The most important deposits in the central area are two iron deposits associated with calc-silicate rocks; the Teltaja deposit with 43 Mt @ 41 % Fe and the Kevus deposit with 38.8 Mt @ 28 % Fe (Frietsch 1985, 1997). In the poorly explored and exposed northern part of the Lannavaara area several smaller iron deposits are found.

The Teltaja iron deposit consists of magnetite and hematite in cherty rocks with minor calc-silicate minerals (Frietsch 1985). The deposit is made up of two mineralised horizons. The first
consists of a 70 m wide magnetite-hematite-bearing body with an average iron content of 47 % Fe, hosted by jaspilitic quartzite. The second is a 15 m wide calc-silicate-bearing magnetite-dominated ore body with average iron content of 33 % Fe. This ore horizon has an anomalously high Mn content (Ambros 1980). The Teltaja mineralisation is almost totally devoid of sulphides. The mineralisation is hosted by supracrustal rocks belonging to the Greenstone Group.

The **Kevus** iron deposit is composed of magnetite with diopside, scapolite and hornblende (Frietsch 1997). The host rock to the mineralisation is impregnated with magnetite. In addition to iron, the mineralisation carries manganese (0.2-0.5 % Mn) and some copper (>0.1 % Cu). The host rock to the mineralisation is metamorphosed, tuffitic basalt. The mineralisation and host rock were highly brecciated and deformed by subsequent tectonic activity.

Another large skarn iron ore is the **Vathanavaara** deposit (51 Mt @ 39.4 % Fe, 0.049 % P and 2.91 % S, Frietsch 1997) hosted by meta-sediments of the Greenstone Group. The ore and host rocks are locally strongly fractured and kaolin-weathered to a depth of at least 100 m. The deposit consists of magnetite with some amphibole. A dark serpentine-bearing ore is also present. The mineralisation is rich in pyrite and pyrrhotite, and small amounts of chalcopyrite occur sporadically. The host rock is a layered graphite-bearing biotite schist. Subordinate to the schist is a scapolite-bearing quartzite with magnetite impregnations.

The **Huornaisenvuoma** deposit is the only Zn-Pb deposit of any economic significance in northern Norrbotten (Bergman et al. 2001). The deposit is hosted by a thick dolomite unit in the upper part of the Greenstone Group (Frietsch et al. 1997). The mineralisation consists of sphalerite, magnetite and pyrite occurring both as disseminated mineralisation and as almost massive layers. The mineralisation generally has a thickness of 1–2 m and its maximum length is 950 m. The country rocks comprise metamorphosed mafic tuff and tuffite, manganiferous iron formation, and black schist. These volcanic and sedimentary rocks were metamorphosed under middle to upper amphibolite facies conditions. Historic ore resource estimates give 0.56 Mt @ 4.8 % Zn, 1.7 % Pb, 0.2 % Cu, and 12 ppm Ag (Frietsch 1991).

The **Kiskamavaara** Cu-Co deposit comprises disseminated and fissure-fill mineralisations of cobalt-bearing pyrite, chalcopyrite, magnetite, hematite and minor amounts of bornite and molybdenite. The host rocks consist of metamorphosed rhyolitic tuffs, intermediate tuffs and mafic volcanic rocks belonging to the Porphyrite Group. Historic ore resource calculations give 3.42 Mt @ 0.37 % Cu and 0.09 % Co at Kiskamavaara (Persson 1981).

The **Kiskamavaara** Cu-Co deposit comprises disseminated and fissure-fill mineralisations of cobalt-bearing pyrite, chalcopyrite, magnetite, hematite and minor amounts of bornite and molybdenite. The host rocks consist of metamorphosed rhyolitic tuffs, intermediate tuffs and mafic volcanic rocks belonging to the Porphyrite Group. Historic ore resource calculations give 3.42 Mt @ 0.37 % Cu and 0.09 % Co at Kiskamavaara (Persson 1981).

**Eastern area**

The most important deposits in the eastern part of the Norrbotten Greenstones are found in a NE-trending, steeply northwest dipping greenstone sequence where the **Tapuli**, **Palotieva** and **Stora Sahavaara** deposits are found (Lindroos 1974 and Table 9).

The **Tapuli** iron deposit is the largest by tonnage in the area and it is the only deposit that has been mined in recent time (2012-2014 by Northland Resources). Similar to the other deposits in the area, the ore body consists of stratiform layers or lenses with a northeastern strike and a dip at 50–65° to the NW. The iron ore is entirely made up of magnetite and skarn minerals, hematite is very rare or absent. Two main types of skarn exist (Lundberg 1967, Lindroos 1972, Lindroos et al. 1972): 1) a serpentine skarn that makes up the gangue or wraps around the iron ore, and 2) a diopside-tremolite skarn that forms a zone between the iron ore and the hanging-wall metasedimentary units, or occurs between the iron ore–serpentine mass and the dolomite. All serpentine is retrograde, replacing tremolite-diopside and all other high-temperature skarns. Most of the ore has accumulated in a fold structure where the ore continues to at least 300 m depth. An important feature of the Tapuli deposit compared to other deposits in the area is the low sulphide content, on average <0.2 % S. The Greenstone sequence hosting the deposits can be followed across the border into Finland.

**Stora Sahavaara** is the third largest deposit in the area (Frietsch 1997, Northland Resources 2007, Northland Resources 2010d). The stratiform deposit forms an arcuate 1300 m long by 40 m thick body that strikes NE and dips 50–70°
Figure 38. Drillcore box from Stora Sahavaara ddh 62001, 63.42 – 77.58 m showing skarn iron ore mineralization. The drillcore log (in Swedish) says:

53.83-75.25 Dark-grey, fine grained, foliated and partly chlorite-altered rock. The last meters with green skarn and decimeter-wide sections with serpentine minerals.

75.25-111.14 Sharp border to fine grained, skarn bearing and partly graphic-bearing magnetite ore (white) with sections of serpentine minerals.

a. Natural colour (RGB) photo
b. False colour composite image in long wave infrared light. Three infrared bands (R=8611 nm, G=10022 nm and B=11810 nm) compose the image and they represent a color rendition of data from sensors measuring across the long wavelength portion of the infrared spectrum. The image has been further processed to a mineral map. Color images and false color composite images of the full core, other cores from Stora Sahavaara as well as other drillcores from Swedish mineralisation can be seen at www.sgu, choose MapViewer; Drill cores.
### Table 9. Iron deposits hosted by Karelian supracrustal rocks (the Greenstone Group) (2.4-1.96 Ga). Data from the Fennoscandian Ore Deposit Database (Eilu et al. 2013)

<table>
<thead>
<tr>
<th>Deposit</th>
<th>When mined</th>
<th>Resources (Mt)</th>
<th>Mined (Mt)</th>
<th>Fe (%)</th>
<th>Mn (%)</th>
<th>P₂O₅ (%)</th>
<th>S (%)</th>
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### Table 10. Base and precious metal deposits hosted by Karelian supracrustal rocks (the Greenstone Group) (2.4-1.96 Ga). Data from the Fennoscandian Ore Deposit Database (Eilu et al. 2013)

<table>
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<th>Deposit</th>
<th>When mined</th>
<th>Resources (Mt)</th>
<th>Mined (Mt)</th>
<th>Cu (%)</th>
<th>Co (%)</th>
<th>Zn (%)</th>
<th>Pb (%)</th>
<th>Fe (%)</th>
<th>S (%)</th>
<th>Ag (g/t)</th>
<th>Au (g/t)</th>
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<tbody>
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<td>35.5</td>
<td>1.7</td>
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<td>Viscaria B-Zon</td>
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<td>Viscaria D-Zon</td>
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<td>Pahtohavare</td>
<td>(1990-1997)</td>
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**Sweden** 309
to the northwest. The bulk of the deposit consists of magnetite and serpentine in diopside-tremolite rocks (Figure 38). There is very little carbonate rock at Stora Sahavaara. Estimates of the ore resources made in the 1960s gave 82 Mt @ 41.0 % Fe, 0.08 % Cu and 2.5 % S to a depth of 435 m (Lundberg 1967). Recent assessment has increased the resource, as shown in Table 9.

Some 15–20 km WSW of Stora Sahavaara is the Pellivuoma iron occurrence. The setting of the Pellivuoma ore bodies, host rocks, ore and gangue minerals, and the relations between the local rock units is similar to that of Tapuli and Stora Sahavaara, except that the deposit is located next to a granite intrusion (Ros et al. 1980; Northland Resources 2010c).

Several smaller skarn iron ores are found in a smaller greenstone area to the west including the Junosuando, Tornefors, Vähävaara and Leppäjoki deposits (Padget 1970, Frietsch 1997). Both skarn iron ores and quartz-banded iron ores are represented in the area. The skarn iron deposits of the Tärendö area are often hosted by carbonate rocks intercalated in the Greenstone Group. They are composed of magnetite which is often Mg-bearing, tremolite, actinolite, diopside, phlogopite, biotite, serpentine and some hornblende. The Ca-Mg-silicates are either evenly distributed throughout the iron ore or occur as layers within the ore. Small amounts of pyrite, pyrrhotite and, locally, chalcopyrite occur.

The quartz-banded iron mineralisations are found at a stratigraphic position similar to the skarn iron ores but are usually made up of quartzites with magnetite and Fe-Mg-Mn silicates; hornblende, grunerite, clinoenstatite, hedenbergite and garnet. The deposits commonly contain some sulfides and minor amounts of manganese. The phosphorous content in both the skarn iron deposits and the quartz-banded deposits is generally very low.

The first deposits in the Tärendö area were already known by 1644 (Frietsch 1997). Between 1846 and 1861 c. 120 tons of iron ore of unknown grade was produced from the Junosuando deposits. None of the other iron deposits in the area have been exploited.

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SELECTED REFERENCES ON THE GEOLOGY OF SWEDEN


Bothnia
Ladozhskoye
Beloye
Onezhskoye
Ozero
Gulf of

Norway
Sweden
Russia
Kiruna
Murmansk
Helsinki

Deposits
- Diamonds
- Hydrothermal fields
- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Size and activity
- Active mine
- Important deposit
- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Scale 1:4 500 000
Stereographic North Pole Projection
Standard Parallel 70° N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 25°E

Gulf of Bothnia

Ladozhskoye Ozero

Kiruna
Murmansk
Helsinki

20° 30°
Archaean

The Archaean crust, either exposed or concealed under Palaeoproterozoic cover rocks and granitoids, occurs in the eastern and northern parts of Finland. The Karelian Province comprises Mesoarchaean 2.8–3.0 Ga lithologies, but rocks older than 3.0 Ga have locally been found. The central part of the Karelian province is mainly Neoarchaean, having plutonic and volcanic rocks of 2.75–2.70 Ga in age. This age difference is also seen in the nature of volcanic rocks, where the older rocks formed in within-plate, probably oceanic, environments, whereas younger volcanic-sedimentary belts show arc-type characteristics (Sorjonen-Ward & Luukkonen 2005, Hölttä et al. 2008). Sanukitoid-type plutonic rocks of the Archaean domain have ages grouping at 2740 and 2718 Ma (Heilimo et al. 2011). The Belomorian province is dominated by 2.9–2.7 G granitoids and includes volcanic rocks formed at 2.88–2.82 Ga, 2.8–2.78 Ga and 2.75–2.66 Ga. The Kola province is a mosaic of Mesoarchaean and Neoarchaean units, together with some Palaeoproterozoic components. The Archaean growth (accretion) of this province occurred from 2.9 Ga to 2.7 Ga and was followed by a collision with the Karelian craton at 2.72 Ga along the Belomorian province.

Palaeoproterozoic cover rocks of the Archaean continents

Rifting of the Archaean crust began in northern Fennoscandia and became widespread after the emplacement of 2.50–2.44 Ga, plume-related, layered gabbro-norite intrusions and dyke swarms (Iljina & Hanski 2005). Erosion and deep weathering after 2.44 Ga was followed by the Huronian glaciation, and later deep chemical weathering again covered large areas in the Karelian province at ca. 2.35 Ga (Laajoki 2005, Melezhik 2006). Rifting events at 2.4–2.1 Ga are associated with mostly tholeiitic mafic dykes and sills, sporadic volcanism and typically fluvial to shallow-water sedimentary rocks (Laajoki 2005, Vuollo & Huhma 2005). Local shallow-marine environments were marked by deposition of carbonates at 2.2–2.1 Ga, showing a large positive δ13C isotope anomaly during the Lomagundi–Jatuli Event (Karlo 2005, Melezhik et al. 2007). 2.05 Ga bimodal felsic-mafic volcanic rocks of alkaline affinity are intercalated with deep-water turbiditic sediments along the present western edge of the Karelian province.

No clear examples of subduction-related magmatism between 2.70 and 2.05 Ga have been found in Finland. The 2.02 Ga felsic volcanic rocks in Lapland (Kittilä) occur in association with oceanic island arc-type rocks and are the oldest candidates for Palaeoproterozoic subduction-related rocks (Hanski & Huhma 2005). Associated continental within-plate volcanic rocks are possibly related to the continuing craton break-up. Bimodal alkaline-tholeiitic magmatism in central Lapland shows that rift magmatism continued further until 1.98 Ga (Hanski et al. 2005). Jormua–Outokumpu ophiolites, tectonically intercalated with deep-water turbidites, are a unique example of Archaean subcontinental lithospheric mantle with a thin veneer of oceanic crust formed at 1.95 Ga along the western edge of the present Karelian province (Peltonen 2005a).

Proterozoic orogenic rocks

The main events in the Palaeoproterozoic orogenic evolution of Fennoscandia can be divided into the Lapland–Kola orogen (1.94–1.86 Ga; Daly
et al. 2006) and the Svecofennian orogen (1.92–1.79 Ga; Lahtinen et al. 2005, 2008). Whereas the Lapland–Kola orogen shows only limited formation of new crust, the composite Svecofennian orogen produced a large volume of Palaeoproterozoic crust in the Svecofennian province.

The Palaeoproterozoic rocks in the Lapland-Kola orogen include small amounts of juvenile, 1.96–1.91 Ga, island arc-type rocks and large volumes of felsic granulites (Daly et al. 2006, Huhma et al. 2011). The oldest rocks in the central Svecofennian province are the 1.93–1.92 Ga island-arc rocks in the Savo belt. Arc-type volcanic rocks are slightly older (ca. 5–10 Ma) than granitoids, but the igneous rocks are predominantly 1.89–1.87 Ga in age in the Central Finland granitoid complex (CFGC) and surrounding belts (Kähkönen 2005). Sedimentary rocks are typically metapsammites with local intercalations of black schists and tholeiitic lavas. Abundant, ca. 1.80 Ga, plutonic rocks occur in northern Finland.
The southern part of Svecofennia in southern Finland (Väisänen & Mänttäri 2002) comprises arc-type volcanism at 1.90–1.88 Ga with partly coeval plutonism at 1.89–1.87 Ga. Sedimentary sequences also include metacarbonate rocks, whereas graphite-bearing rocks are rare. Two metamorphic peaks, at 1.88–1.87 and 1.83–1.80 Ga, have been detected in southern Svecofennia. A major unconformity between them is indicated by the occurrence of lateritic palaeosols (Lahtinen & Nironen 2010) and ≤ 1.87 Ga quartzites and meta-arkoses (e.g., Bergman et al. 2008). Younger syn- to post-tectonic granites (1.85–1.79 Ga) are common in southern Svecofennia.

Rocks of the Mesoproterozoic rapakivi granite association (1.65–1.47 Ga) are locally voluminous and especially characteristic for southern Finland (Rämö & Haapala 2005).

MINING HISTORY - FINLAND
Pasi Eilu (GTK)

16th – 19th Centuries
The Ojamo iron ore mine, which started production in 1530, is regarded as the first metal mine in Finland. Following this, over 350 metal mines had been in operation before World War II. The scale of production in these mines was modest, although mining played an important role in the slowly developing society. Before the 1920s, the mines mainly produced iron ore for iron works in southern Finland. Sulphide ore production was mostly from one mine, Orijärvi (Cu-Zn) in SW Finland. From 1530 until the end of the 19th Century, metal ore production totalled 1.4 Mt, of which sulphide mines comprised 1.0 Mt (most of which was produced after 1850) and the iron mines 0.4 Mt (Puustinen 2003).

20th Century
In Finland, the modern mining industry started to form along with the Outokumpu mine. The deposit was discovered in 1910 (Stigzelius 1987) and in 1928 it became the largest sulphide ore mine in the country. Small-scale production started immediately in 1910, production gradually increased in the 1920s and 1930s and total ore output was almost 6 Mt between 1930 and 1945. During its lifetime, 1910–1989, about 28 Mt of ore was mined and 1 Mt copper produced (Puustinen 2003). The Petsamo (Pechenga) nickel deposit was found in 1921, in the then north-easternmost corner of Finland (Haapala & Papunen 2015). The development of a mine at Petsamo was complicated, but eventually, in the period 1936–1944, about 0.5 Mt of ore was mined, first as a Finnish-Canadian cooperation, and during World War II by Germany. The war between Finland and the Soviet Union ended in September 1944, and the Petsamo region was subsequently ceded to the Soviet Union.

Soon after the war, in the late 1940s, the Aijala (Cu) and Otanmäki (Fe-Ti-V) mines were opened. Otanmäki gradually developed into a globally significant vanadium mine, responsible for about 10 % of the world’s vanadium production during the 1960s and 1970s (Ilí et al. 1985). Seven metal mines were opened in the 1950s, including the Vihtanti (Zn) and Kotalahti (Ni) mines. The most active mine development period in Finland was 1960–1980, when more than twenty metal mines started production. The most important were the still operating Kemi Cr and Pyhäsalmi Zn-Cu mines. As a consequence, total yearly metal ore output peaked in 1979 at slightly over 10 Mt. A few small mines were opened in the 1980s, but a number of major mines were closed in the same period, and total production gradually declined to about 3 Mt in the early 2000s (Puustinen 2003).

Before the opening of the Talvivaara and Kevitsa mines in 2008 and 2012, respectively, the...
largest sulphide mine in Finland was Pyhäsalmi. The deposit was discovered in 1958 when a local farmer dug a well through the overburden till into a subcrop of the massive ore (Helovuori 1979). By the end of 2014, over 53 Mt of ore had been mined and the remaining ore has secured a further 5 years of production.

The Kemi chromite deposit was found 1959 by a local layman (Alapieti et al. 2005). Open-pit chromite mining began in 1966 and ferrochrome production in 1967 at nearby Tornio, at the far northern end of the Gulf of Bothnia. Stainless steel production at Tornio commenced in 1976. In 2006, the underground mine became the sole source of ore. Its current design capacity is 2.7 Mt/y of ore. The known ore reserves will enable mining to continue for several decades, and a recent seismic reflection survey suggests that the ore extends possibly beyond 2 km depth (Huhtelin 2015).

Currently, we are living in a new era in Finnish mining history. Two major mines, Kittilä Suurikuusikko (gold) and Talvivaara (nickel), were opened in 2008, and the Kevitsa Ni-Cu-PGE mine in 2012. These three mines have multiplied Finnish metal ore output to about 14 Mt/y in 2014. In addition to these major deposits, a number of smaller projects have recently started; these include the Jokisivu mine which produced its first gold in 2008, and the Pampalo gold and the Kylylahti Outokumpu-type Cu-Co-Ni-Zn mines which opened in 2011. The Laiva (Laivakangas) gold mine in western Finland also went into production in 2011, but is presently in care and maintenance.

The Talvivaara Ni-Zn-Cu-Co deposit was discovered in 1996 (Loukola-Ruskeeniemi & Heino 1996). The resource was found to be large but of low grade, and it was concluded at the time that exploitation was not economically viable using conventional metal extraction techniques. Bioheap leaching was later found to be a suitable method for operating this sulphide deposit. The mine successfully produced the first metals in October 2008, and has been in production since 2010. With the total resource of 2100 Mt @ 0.22 % Ni, 0.50 % Zn, 0.13 % Cu and 0.02 % Co, Talvivaara has the potential to become a globally significant producer, especially for nickel. However, the bio-heap leach process, a global first for a Ni mine, and the complete metal extraction process, have had a number of complications, and the mine has yet to reach the expected production levels.

The Geological Survey of Finland (GTK) discovered gold in the Suurikuusikko area of the Central Lapland greenstone belt in 1986 (Wyche et al. 2015). A preliminary estimate of inferred resources in 2000 amounted to 8.3 Mt @ 6.1 g/t gold. Construction of the mine began in June 2006 (Agnico Eagle Mines Ltd), and the mine was named after the local municipality of Kittilä. The first gold was poured on 14 January 2009. The annual production has been 4-5 tons of gold, but is expected to increase to 7.5 tons in 2016. At the end of 2014, the total pre-mining resource (total resources + mined ore) amounted to 57.6 Mt @ 4.38 g/t Au. The deposit is known for about 4 km along strike, and is open both along strike and to a depth of 1.5 km.

The Kevitsa Ni-Cu-PGE deposit was found in 1987 (Mutanen 1997). Gold Prospecting AB (later Scandinavian Minerals Ltd.) claimed the deposit in 2000 and started planning for a mine. First Quantum Minerals Ltd. (FQM) bought Scandinavian Minerals Ltd. in 2008, built the mine, and started production in 2012. The current pre-mining resource (total resources + mined ore) is 272 Mt @ 0.40 % Cu, 0.30 % Ni, 0.015 % Co, 0.24 g/t Pd, 0.19 g/t Pt, and 0.11 g/t Au. In 2014, 7 million tonnes of ore was processed at Kevitsa.

By far the largest industrial minerals mine in Finland is the Siilinjärvi mine (Kemira Corp. 1979–2007, Yara International ASA 2007–), hosted by an Archaean carbonatite intrusion (O’Brien et al. 2015). First indications of the carbonatite were found in a railway cutting in 1950, and exploration began in full by GTK in 1958. The deposit was test-mined during 1975–1979, and full-scale production started in 1980. In 2014, the mine produced 11 Mt ofapatite ore, from which about 0.95 Mt tonnes ofapatite concentrate were recovered as the main product, with by-product carbonates and mica also recovered. By the end of 2014, 271 Mt of ore has been mined at Siilinjärvi: the remaining resources are reported to be 1617 Mt @ 3.8 % \( P_2O_5 \).
PYHÄSALMI AND VIHANTI DEPOSITS
Jouni Luukas & Kaj Västi (GTK)

The Vihanti–Pyhäsalmi area is located on the NE edge of the Svecofennian domain, in the NW part of the Raahe–Ladoga suture, previously commonly called the Main Sulphide Ore Belt of Finland (Kahma 1973). The Raahe–Ladoga suture has been described as a collisional boundary zone between Proterozoic and Archaean domains (Lahtinen 1994). The Vihanti–Pyhäsalmi area comprises the central part of the northwestern Savo schist belt (Vaasjoki et al. 2005) and is 10–40 km wide and about 300 km long.

Most of the massive sulphide deposits in the area are hosted by metavolcanic rocks, locally also by metasedimentary rocks, in a Palaeoproterozoic island arc environment, close to the Archaean Karelian craton. In the Pyhäsalmi region, volcanic activity started in an extensional continental margin with felsic volcanism and continued in a rifted marine environment with mafic volcanism. Large-scale hydrothermal alteration and mineralisation occurred close to the centres of mafic volcanism. This 1.93-1.92 Ga bimodal volcanic event represents the lowermost volcanic unit in the Raahe-Ladoga Zone. In the northwestern end of the area (Vihanti area), volcanogenic rocks are predominantly intermediate and felsic in composition. These compositionally different units have been named as the Pyhäsalmi and Vihanti groups. Without a longer hiatus, volcanic activity continued at 1.89–1.88 Ga with more calc-alkaline volcanism (Kousa et al. 1997). The Vihanti–Pyhäsalmi metallogenic area includes a large number of Zn-Cu deposits and prospects. It is particularly known for the two world-class VMS-type Zn deposits at Pyhäsalmi and Vihanti, but there also are a few smaller mines and a number of unexploited deposits and occurrences. Currently, only the Pyhäsalmi mine is active. In the vicinity of Pyhäsalmi there are three smaller mined VMS-type deposits: Mullikkoräme, Ruostesuo and Kangasjärvi.

Pyhäsalmi

The metavolcanic rocks of the Pyhäsalmi group in the Pyhäsalmi mine area belong to the Ruotanen Formation (Puustjärvi 1999). Voluminous piles of sodium-rich quartz plagioclase-phryric rhyolitic metavolcanic rocks with abundant mafic dikes form the lowermost part of the formation. This Kettuperä gneiss unit (sample A0751 in Helovuori 1979) in the Pyhäsalmi group is interpreted here as the oldest felsic volcanic member of the Ruotanen Formation. New zircon U-Pb dating results give an age of 1924±3 Ma, which is a reliable age for volcanism in the Pyhäsalmi group (Kousa et al. 2013). This unit is overlain by voluminous felsic pyroclastic breccias which are usually totally altered into cordierite-sericite schists. The pyroclastic unit shows abundant sulphide dissemination and the major sulphide ore is located here. The felsic volcanic rocks are overlain by massive mafic lavas which form the uppermost part of the Ruotanen Formation. The intermediate volcanic rocks with minor calc-silicate interlayers belong to the overlying Vihanti group.

The Pyhäsalmi deposit is hosted by a volcanic sequence composed of lapilli tuffs, coherent lava flows and sill-shaped intrusions of varied composition (Figure 2). Rhyolitic volcaniclastic rocks are the most common host rock near the sulphide deposit. Total production since the beginning of mining has been 50 Mt (end of 2012) grading 0.92 % Cu, 2.47 % Zn and 37.2 % S. The ore contains, on average, 0.4 g/t Au and 14 g/t Ag (Mäki et al. 2015).

Vihanti

Volcanic activity in the Vihanti area took place in two stages (Rauhamäki et al. 1980). The earlier cycle started under marine conditions and com-
prised felsic volcanic rocks and chemically precipitated carbonate rocks, and is characterised by the formation of Lampinsaari-type ore bodies. The deposition of volcanic and carbonate rocks continued concurrently with ore formation, culminating in the stages of zinc ore deposition.

After mineralisation, more felsic volcanic rocks were erupted during the later cycle. Graphite tuff between the zinc ore and the later volcanic cycle refers to reducing conditions (op. cit.).

The Vihanti Zn-Pb-Ag deposit is situated in the northwestern part of Raahe-Ladoga zone. An essential part of this rock assemblage is composed of what is called the Lampinsaari-type rock association comprising felsic metavolcanic rocks, calc-silicate rocks and graphite tuffs (Laukas et al. 1998).
The deposit consists of five types of mineralisation: zinc, copper, pyrite, lead-silver-gold and uranium-phosphorous ore (Figure 3). According to Autere et al. (1991), the U-P type is the oldest, whereas the Pb-Ag and Zn ores are the youngest. There are about 20 separate ore bodies that have been mined out. The metal content and size varies greatly between the different bodies. The Zn ores, hosted by dolomite and calc-silicate rocks, were by far the most important ore types for the economy of the mine, containing over 75% of the total ore resource. Minor Zn ore bodies were mined out along the southern fold limb down to +800 level. Zn-ores are typically situated in tectonically thickened skarn-dolomite-serpentinite beds in the upper levels of the deposit. Although the Zn-ores contain some Ag and Au, a separate disseminated Pb-Ag ore type is located in close connection to the Zn-ores. The best Pb-Ag ores were in the western part of the mine. Pyrite- or pyrrhotite-rich ores which are situated in the upper parts of the mineralised horizon are called pyrite ores. The Hautaräme and Hautakangas ore bodies on the downward dipping limb area are compact pyrite ores hosted by felsic volcanic rocks and calc-silicate rocks. Cu-ores in the felsic volcanic rocks are disseminated ore types, which are closely related to the pyrite ores. The fifth separate ore type is uranium-phosphorous ore which forms a non-continuous layer in the upper part of U-P-horizon, the downward dipping limb area. This unexploited ore type is hosted by felsic volcanic rocks interbanded with calc-silicate rocks (Rehtijärvi et al. 1979). The underground production from the Vihanti mine was 28 Mt ore which contained 4% Zn, 0.4% Cu, 0.36% Pb, 25 ppm Ag and 0.44 ppm Au. The U-P mineralisation, with over 1 Mt of low grade ore, has never been exploited. In addition to the Vihanti deposit, there are a couple of smaller unexploited deposits. Although intense exploration has been carried out, only minor showings have so far been detected in the Vihanti district.
The Outokumpu deposit, which was mined until 1984, contained c. 30 million ton @ 3.8 wt.% Cu, 0.24 wt.% Co, 1.07 wt.% Zn, 0.12 wt.% Ni, 0.8 g/t Au and 8.9 g/t Ag. Since the discovery of the Outokumpu ore in 1910 a dozen more similar, but smaller and metal-poorer, semimassive to massive Cu-Co-Zn sulphide deposits have been discovered in the Outokumpu mining district. Three of these have been exploited, including Vuonos, Luikonlahti and the currently mined Kylylahti deposit (Peltonen et al. 2008; Kontinen 2012a). The host assemblage of the Outokumpu deposits also contains low-grade Ni occurrences that were mined at Vuonos for 5.5 million ton @ 0.2 wt.% Ni.

The Outokumpu mining district is part of the Palaeoproterozoic North Karelia Schist Belt and is confined within the boundaries of the Outokumpu Allochthon (also known as the ‘Outokumpu Nappe Complex’ or ‘Outokumpu Nappe’). It contains serpentinite bodies fringed with the dolomite-skarn-quartz rock and black schist host rock assemblage of the polymetallic and polygenetic Outokumpu-type copper-cobalt-zinc and Kokka-type nickel mineralisation.

More than 85 % of the Outokumpu Allochthon is composed of schistose, metaturbiditic wackes and pelites, which occur in sequences dominated by medium to thinly bedded psammites. Metavolcanic rock intercalations and synsedimentary magmatic intrusions are not found in the apparently kilometres-thick, now complexly folded and faulted, turbidite package. Models involving relatively deep water deposition of the turbidites, probably at continental slope-rise fans in a passive margin type environment, have been proposed (Kontinen & Sorjonen-Ward 1991, Peltonen et al. 2008, Lahtinen et al. 2010). The wacke schist sequence contains layers of black schists (partly black turbidite muds), which appear to be thicker and more common in the presumably lowermost parts of the allochthon (Loukola-Ruskeeniemi 1999, Kontinen et al. 2006). These schists contain abundant organogenic graphite (5–10 %) and commonly also iron sulphides (5–20 wt.%). They are strongly enriched in Ni, Cu and Zn relative to the upper crust or average shale compositions (Loukola-Ruskeeniemi 2011). The cobalt content is systematically relatively low, 33 ppm on average. The low Co distinguishes the black shales from the semimassive to massive Co-Cu-Zn mineralisation, which has much higher Co contents and Co/Ni ratios. The black-schist sulphides also have isotope compositions indicating a crustal source of their Pb, versus the obvious mantle source of Pb in the most pristine parts of the sulphide bodies (Peltonen et al. 2008). Redox-sensitive metals such as Sb, Se, Mo, V and U occur in the black schists in elevated concentrations (Kontinen et al. 2006; Loukola-Ruskeeniemi 2011), attesting to at least periodically anoxic-sulphidic depositional conditions, but also to a generally oxygenated atmosphere at the time of the Outokumpu sedimentation (Kontinen & Hanski 2015).

The turbiditic wacke-shale sediments in the Outokumpu Allochthon were deposited subsequent to 1920±20 Ma, which is the age of their youngest dated detrital zircon grains (Lahtinen et al. 2010). The black-schist interleaved lower units of the allochthon host numerous fault-bound (exotic) bodies of mantle peridotite-derived serpentinite with variable components of 1.96 Ga gabbroic, basaltic and plagiogranitic rocks (Peltonen et al. 2008). Because of their older age than the sedimentation of the enclosing turbidites, these peridotite-gabbro-basalt bodies are interpreted as fragments of 1.95 Ga old mafic-ultramafic oceanic floor/crust, tectonically incorporated into the turbidite sediments during the early obduction of the Outokumpu Allochthon at ca. 1.90
Ga (Peltonen et al. 2008). Serpentinites in the ophiolite fragments show at their margins, especially where located against graphitic-sulphidic schists, an omnipresent listwaenite (carbonate, carbonate-silica) to birbirite (silica) alteration. Where ideally developed, the alteration zonation is, from serpentinite to black schist: 1–5 m of carbonate, 5–15 m of carbonate-silica (listwaenite), and 15–40 m of silica rock (birbirite). These zones are metamorphosed and recrystallised to dominantly carbonate, tremolite±diopside and quartz rocks, respectively. Attesting to their peridotitic origin all these rocks consistently contain 1000–4000 ppm Cr and Ni, and nearly always chromite. Particularly thick alteration zones are typically flanked by especially thick zones of sulphide-rich black schist.

Some of the Outokumpu area serpentinite bodies host, in their deformed-metamorphosed carbonate-silica alteration envelopes, massive to semi-massive Cu-Co-Zn and disseminated Ni sulphides of the Outokumpu and Kokka type, respectively. Of these two mineralisation types, the economically far more significant Outokumpu Co-Cu-Zn is, in all cases, found with a parallel zone of Ni-dominated mineralisation in the hosting skarn-quartz rocks. In contrast, the Kokka Ni mineralisation occurs widely also in environments without any Co-Cu-Zn mineralisation. Weak Ni mineralisation is, in fact, a ubiquitous feature of the Outokumpu alteration assemblage, which systematically contains 1000–4000 ppm whole rock Ni predominantly located in disseminated pyrrhotite and pentlandite. The concept of Kokka-type Ni mineralisation refers to irregular zones of more elevated Ni contents and Ni sulphides within the alteration zones, commonly in locations of intense shearing and in proximity to highly sulphidic black schists (Huhma 1975).

Within the Outokumpu district there are only minor differences between the individual deposits apart from size and features caused by variations in metamorphic grade, which increases in the area from east to west, from lower amphibolite to upper amphibolite facies (Säntti et al. 2006; Kontinen et al. 2006, Peltonen et al. 2008). All the ore bodies contain, as their main sulphide minerals, variable contents of pyrite, pyrrhotite, chalcopyrite, sphalerite, pentlandite and locally cobaltite. The dominant gangue mineral is usually quartz with some tremolite±diopside although locally the latter may predominate. The sulphide bodies at Kylylahti (Figures 4 and 5), where the metamorphic grade is lowest, are predominantly pyritic and show elevated contents of more ‘volatile’ elements such as As, Sb and Hg, whereas the ore bodies in upper amphibolite/migmatite grade environments are generally pyrrhotitic with much lower contents of Sb, As and Hg. The impact of the metamorphic control (cf., Vaasjoki et al. 1974) also results in the presence of a large part of the Co content being in pyrite at Kylylahti, whereas at Luikonlahti nearly all the cobalt is in pentlandite. The large Outokumpu deposit is partly pyritic and relatively rich in As and Au, whereas the thinner, smaller Vuonos and Perttilahti deposits are dominantly pyrrhotitic and very low in As and Au. The effects of metamorphism on the Kokka type Ni occurrences are restricted mainly to an increase in the grain size of sulphides and gangue and increased remobilisation of the sulphides into blotches and veinlets.

The Outokumpu Co-Cu-Zn sulphides are interpreted to be polygenetic with a primary inhalative-exhalative Cu-Co-Zn proto-mineralisation event(s) at ca. 1.95 Ga, in a hot spring environment in spreading zones or leaky transforms in a predominantly ultramafic ocean floor (Peltonen et al. 2008). The Ni mineralisation of the carbonate-silica alteration assemblage occurred during the ca. 1.90 Ga obduction of the ophiolite fragments, inside the host turbidite sediments, when low T (<200ºC) carbonaceous and sulphidic fluids altered the outer margins of the ultramafic bodies to sulphide-bearing carbonate - quartz rocks. Nickel originally bound in ferromagnesian mantle minerals, such as olivine, was relocated into sulphides during this process. The exceptionally high Ni content in the Co-Cu type ores is interpreted to reflect early syntectonic interaction of the Cu-Co-Zn protosulphides with the Ni sulphides, by simple ‘mechanical’ mixing of the two sulphide end-members or by fluid-assisted diffusion of Ni from the Ni occurrences into the Cu-Co bodies. Later syntectonic solid-state remobilisation and concentration of the Cu-Co-Zn-Ni sulphides into late fault-controlled positions completed the geometric style.
of the Outokumpu type deposits as thin (1–15 m) and narrow (50–450 m) but long (1–>5 km) sheets. In such serpentinite bodies or ophiolite fragments that obducted without any Cu-Co protosulphides, only irregular local Ni enrichments of the Kokka type were generated in structurally favourable positions.

In exploration for Outokumpu-type ores, it is important to recognize that valuable sulphides occur only in close proximity to serpentinite bodies of which only a fraction are found to be mineralised, obviously those that were obducted with Cu-Co protosulphides. Although some authors emphasise the role of black schists (e.g. Loukola-Ruskeeniemi 1999), it should be noted.
that on the basis of Pb isotope compositions, the commonly accompanying black shales did not contribute significantly to the genesis and metal budgets of the ore bodies (Peltonen et al. 2008). Experience from past exploration largely confirms this inference as such characteristics as metal contents and ratios of the black schists have nowhere been observed as useful vectors to the Cu-Co ores. The genesis of the Outokumpu alteration assemblage and of the Kokka-type Ni occurrences did, however, involve the influx of S and of certain metals, such as As, Sb and Pb, from the black shales.
The Talvivaara deposit, discovered by the Geological Survey of Finland in 1977, is a very large (2 billion metric tons) low-grade, black-shale hosted, polymetallic sulphidic resource containing 0.23 wt.% Ni, 0.54 wt.% Zn, 0.13 wt.% Cu, 0.02 wt.% Co and also 16 ppm recoverable U (Loukola-Ruskeeniemi & Lahtinen 2013; Kontinen & Hanski 2015). The deposit, which has been intermittently mined since 2007, is currently the largest sulphidic Ni resource under exploitation in Western Europe. Talvivaara is located in central Finland within the Palaeoproterozoic, dominantly metasedimentary, Kainuu Schist Belt (Laajoki 1991, 2005). Several similar but much smaller mineralisations have been observed elsewhere in the central and southern part of the Kainuu belt and a couple also further south in the northernmost part of the North Karelia Schist Belt (Kontinen 2012b). Talvivaara and the smaller deposits have been found on the basis of ground and low-altitude airborne geophysical and moraine- and lithogeochemical surveys, with follow-up including mapping, boulder tracking and drilling operations.

The main components of the Proterozoic strata in the Kainuu Schist Belt (Figure 6) are: (1) Sumi-Sariola and Jatuli stage, 2.5–2.1 Ga, cratonic and epicratonic, dominantly feldspathic to quartz arenite sequences, (2) Lower Kaleva stage, 2.1–1.95 Ga, rift-related wacke-pelite sequences, and (3) Upper Kaleva stage, 1.95–1.90 Ga, deep-water turbidite wackes and pelites (Kontinen 1986, 1987, Laajoki 2005). Of these sequences, the two older ones are autochthonous strata on a deeply eroded, mainly late Archaean, mostly gneissic-granitoid basement, whereas the third is at least partly in allochthonous units carried by fault-bound, exotic ophiolitic fragments of 1.95 Ga oceanic crust (Peltonen et al. 2008 and references therein). The Lower and Upper Kaleva differ for their sediment sources, which were dominantly Archaean and/or recycled Archaean for the former and Proterozoic (mostly 1.92-2.0 Ga) for the latter (Kontinen & Hanski 2015).

The Talvivaara-type deposits occur in association with sulphide- and metal-rich carbonaceous (now graphitic) sediments. Carbonaceous sediments first appear as thin layers in the topmost 2.1 Ga dolomitic and tuffaceous strata of the Jatuli sequence. They become abundant in the upper parts of the Lower Kaleva sequence, after ca. 2000 Ma (Kontinen & Hanski 2015). Layers of carbon- and sulphide-rich sediments are also common in the deeper-water feldspathic wacke turbidites of the Upper Kaleva stage.

The giant Talvivaara deposit has an exposed strike length of about 12 km (Figure 6). It comprises one or two, originally probably <50 m thick layers of strongly metal-enriched, thick bedded to laminated graphite- and sulphide-rich muds. Muds are intercalated with cm to meter thick layers of thinly bedded to laminar pyritic muds and carbonate rocks. The primary muds have been pervasively metamorphosed and recrystallised into more coarse-grained, often quartz-plagioclase-microcline veinined, quartz-anorthite-biotite-muscovite gneisses/chists interbedded with carbonate-diopside-tremolite calc-silicate rocks (Ervamaa & Heino 1980, Loukola-Ruskeeniemi & Heino 1996, Kontinen & Hanski 2015).

Based on drill core observations, footwall contacts the ore-bearing unit (Talvivaara Formation) grade, with the appearance of quartz wacke interbeds, into a quartz wacke unit at least hundreds of metres thick containing variably abundant interlayers of graphitic-sulphidic phyllite, meta-carbonate rocks and mass-flow conglomerate (Hakonen Formation) (Figure 6, 7). The hangingwall contact of the ore involves a rapid shift to black shale-intercalated, graphite-rich felds-
pathic wackes (Kuikkalampi Formation), representing a change, in terms of sediment source, to Upper Kaleva deposition. In its middle part, the Talvivaara deposit is overlain by a small klippe of the allochthonous Upper Kaleva turbidites with thin fault-bound lenses of tale-carbonate altered ophiolitic mantle peridotites spread all along its basal contact.

Upright compressional folding and related reverse faulting in a late stage of the tectonic deformation have contributed significantly to the volumes of minable ores by tectonic thickening of the mineralised unit. The present sulphide mineral assemblage is pyrrhotite-pyrite-

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\text{sphalerite-chalcopyrite-pentlandite±alabandite, with a high variation in the pyrrhotite-pyrite ratio. All the sulphides have undergone metamorphic equilibration involving Fe derived from the epiclastic detritus; only part of the pyrite in the samples, which were originally rich in syngenetic-dia-genetic pyrite, seems to have avoided complete re-equilibration, which involved generation, mobilisation and concentration of monosulphide solid solution (mss) and intermediate solid solution (iss) in syntectonic veinlets and blebs (Kontinen et al. 2013b; Kontinen & Hanski 2015). As a consequence all the main sulphides, except pyrite, have near constant compositions throughout the minerali-}
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sation; e.g. pyrrhotite has 4930±690 ppm Ni, sphalerite 8.1±1.0 wt.% Fe and pentlandite <100 ppm Co. Besides sulphides there is significant iron only in the phlogopitic biotite. Metamorphic processes were important for the utilisation of the deposit as they relocated Ni in the easily soluble pyrrhotite and pentlandite from precursor Ni-rich phases which may have comprised high-Ni pyrite, millerite, polydymite or perhaps organometals. Pyrite contains the main part of the Co in the deposit, predominantly in coarse metamorphic pyrite (Kontinen & Hanski 2015).

In all the known Talvivaara-type occurrences the mineralised unit is characterised, for its best parts, by distinctly graphite-rich (5-15 % C as graphite) and sulphide-rich (5-30 % S in iron sulphides) mud-dominated beds with calc-silicate rock and quartz wacke intercalations. Mineralised units are located, in several cases immediately above phosphorite-chert-black shale sequences intercalated with silicate- and oxide-facies iron formations. In the largest of the known deposits, Talvivaara, the whole mineralised unit is, over its whole known extent, surprisingly uniform in its metal content and ratios. All the known Talvivaara mineralisations show a similar black-shale type enrichment in C, Fe, S and redox-sensitive metals compared to the background ‘barren’ black shales. Mineralised muds also show Co/Ni, Cu/Ni or Zn/Ni ratios broadly similar to the accompanying ‘barren’ black schists, or Kalevian black shales in general.

Following the discovery of the Talvivaara deposit, a submarine hydrothermal source of the excess base metals was proposed (Ervamaa & Heino 1980), and has remained the explanation, with some variations, in several subsequent studies (Loukola-Ruskeeniemi 1991, Loukola-Ruskeeniemi & Heino 1996, Loukola-Ruskeeniemi 1999). Recently, Kontinen (2012b) has presented, and Kontinen et al. (2013c) and Kontinen & Hanski (2015) further refined, an interpretation of the Talvivaara mineralisation as a synsedimentary metal enrichment under a water column that was apparently distinctly enriched in Ni, Co, Cu and Zn. The same mechanisms are most likely responsible for producing the metal enrichment in the associated ‘barren’ graphitic metasedimentary units. The elevated Mn and Fe contents provide support for effective cycling of Mn and Fe hydroxides-oxyhydroxides in basinal redoxclines also contributing to the high base metal levels. The very high Fe (±Mn), S and base metal contents (for black shales), as at Talvivaara, probably require far-field (as no near-field are present) hydrothermal sources. The uniformly high Ni/Cu (1.9 ± 0.6) and Ni/Zn (0.5 ± 0.1) ratios provide, however, evidence for the derivation of the base metals from a well-mixed and very large, probably oceanic seawater-type reservoir. The common presence of up to metre-thick, compositionally highly monotonous massive muds in the mineralised units as in Talvivaara implies an important role for intrabasinal resedimentation-recycling. Local cm to meter scale post-depositional metal redistribution has probably taken place to some extent already during diagenesis and early metamorphism but certainly during peak-metamorphism in the form of structurally controlled solid-state mss and iss migration with concentration into veinlets and blebs that characterize the more structurally reworked parts of the Talvivaara deposits.
Suurikuusikko is by far the largest of the many epigenetic gold (±copper) deposits of the Central Lapland greenstone belt. The Central Lapland greenstone belt (CLGB) is the largest mafic volcanic-dominated province preserved in Fennoscandia. It extends from the Norwegian border in the northwest to the Russian border in the southeast (and has continuations in each of the neighbouring countries). The belt was initially formed in an intracratonic rift setting related to the breakup of the Archaean Karelian craton (Hanski & Huhma 2005), and filled by clastic and chemical sedimentary and rift-related volcanic rocks, and in the central parts, by juvenile oceanic basaltic and komatitic volcanic rocks. Extensive, apparently early, albitionisation and locally abundant scapolite in amphibolite-facies parts of the region suggest that evaporites may also have been deposited in this rift environment. The area is characterised by several phases of mafic to ultramafic intrusive activity, and on the top of the sequence, by molassic sedimentary units. The deformational evolution of the CLGB in the period 1.92–1.77 Ga took place in four to five major stages: D₁ and D₄ during 1.92–1.88 Ga, D₃ ± D₂ either during 1.92–1.88 Ga or 1.84–1.79, and the latest stage (D₄ or D₅) at ca. 1.77 Ga or slightly post-1.77 Ga (Hanski & Huhma 2005, Lahtinen et al. 2005, Patison 2007, Saalmann & Niiranen 2010, Lahtinen et al. 2015).

The Kittilä metallogenic area covers most of the CLGB. Its extent is essentially defined by various indications of orogenic gold mineralisation with about 40 drilling-indicated deposits and occurrences currently known. These include Suurikuusikko, the largest presently active gold mine in Europe (the Kittilä Mine), one closed mine (Saattopora), and several test-mined deposits. Orogenic gold mineralisation within the CLGB can be further divided into gold-only and anomalous metal-association subtypes, as these categories were originally defined by Groves (1993) and Goldfarb et al. (2001), respectively. Within the CLGB, both subtypes are characterised by sitting in lower-order structures, Au/Ag >1, quartz veining is abundant, the sulphide contents are 1–10 vol%, the dominant ore minerals are pyrite, arsenopyrite and pyrite, carbonisation and sericitisation haloes surrounding the mineralisation (Eilu et al. 2007, Eilu 2015). Another significant factor in locally controlling gold mineralisation within the CLGB, in addition to structure, is the pre-mineral albitionisation which prepared the ground by creating locally most competent units, which thus provided the best sites for local dilation, veining and mineralisation (Saalmann & Niiranen 2010). Many of the occurrences of the Kittilä metallogenic area are of the gold-only style, including Suurikuusikko, but more than a half of the occurrences also contain Co, Cu and/ or Ni as potential commodities, and can be included into the subcategory of ‘anomalous metal association’ as defined by Goldfarb et al. (2001).

The Suurikuusikko deposit had a pre-mining gold endowment of more than 8 million ounces. The deposit is hosted by tholeiitic mafic volcanic rocks of the ca. 2.02 Ga Kittilä Group of the CLGB (Figure 8). The gold is refractory, occurring in arsenopyrite and pyrite, and the mineralisation is associated with intense pre-gold albite and syn-gold carbonate-sericite alteration. The deposit is within the sub-vertical to steeply east-dipping Kiistala shear zone. Numerous ore lenses are distributed along, and within this N-to NNE-trending structure (Figure 8). Individual, subparallel ore lenses appear to be controlled by multiple phases of deformation and generally have a moderate northerly plunge. Mineralised intervals have been found for over 5 km strike length in the host structure, and from the surface to a depth of >1.6 km. Four stages of sulphide formation have been detected at Suurikuusikko.
Gold is associated with the second stage of arsenopyrite and pyrite growth. A Re–Os age of 1916±19 Ma has been obtained from gold-bearing arsenopyrite. This suggests that mineralisation took place 60–100 Ma after Kittilä Group deposition and before the end of collision-related sedimentation in the CLGB. This age for Suurikuusikko is similar to that of the SW-directed thrusting event related to the CLGB. Suurikuusikko has nearly all the features typical for an orogenic gold deposit. The relatively early apparent timing of this deposit in the orogenic evolution of the CLGB, albited host rocks, fine-grained carbon in the host rocks and auriferous veins, and the dominance of refractory gold are less commonly documented features in orogenic gold deposits, but do not suggest an alternative genetic type for this deposit and are not inconsistent with an orogenic gold system. (Patison et al. 2007, Koppström 2012, Wyche et al. 2015).

Figure 8. Geological map of the Suurikuusikko area.
The Hannukainen Fe(-Cu-Au) deposit located in the Kolari district, western Finnish Lapland, is the largest known magnetite deposit in Finland. About 10 similar iron deposits are known from the Kolari district which forms the westernmost part of the Central Lapland greenstone belt (CLGB). The CLGB was formed during multiple stages of rifting with deposition of volcanic and sedimentary rocks in intracratonic and cratonic margin settings between 2.5 and 2.0 Ga and was subjected to metamorphism and deformation during Svecofennian orogenic events between 1.91 and 1.77 Ga (Lehtonen et al. 1998, Hanski & Huhma 2005). The supracrustal rocks of the Kolari region consist of ≥2.2 Ga Sodankylä Group quartzites and overlying 2.05 Ga Savukoski Group mafic volcanic rocks, mica schists and gneiss, dolomitic marbles and graphite bearing schists. These are overlain by <1.89 Ga Kumpu Group quartzites, conglomerates and mica schists (Hiltunen 1982, Väänänen & Lehtonen 2001). Intrusives in the area consist of 1.86 Ga Haparanda Suite monzonites and diorites and ca. 1.80 and 1.76 Ga granites (Hiltunen 1982, Niiranen et al. 2007). Structurally the Kolari district is located at the Baltic-Bothnian megashear (BBMS) which has been interpreted to represent a cratonic boundary between the Norrbotten and Karelian cratons to the west and east, respectively (Berthelsen & Marker 1986, Lahtinen et al. 2015). The dominant host rocks for the known deposits are the altered Savukoski Group mafic volcanic rock or 1.86 Ga diorite or a combination of these (Hiltunen 1982, Niiranen et al. 2007). Structurally the Kolari district is located at the Baltic-Bothnian megashear (BBMS) which has been interpreted to represent a cratonic boundary between the Norrbotten and Karelian cratons to the west and east, respectively (Berthelsen & Marker 1986, Lahtinen et al. 2015).

The deposits display strong structural control by the shear and thrust structures comprising the BBMS in the area (Niiranen et al. 2007, Moilanen & Peltonen 2015).

Mineralisation at Hannukainen took place in two stages; the initial magnetite-stage was preceded by a Cu-Au stage. Fluid inclusion data suggest...
that the mineralizing fluids were moderately saline (32-56 wt.% NaCl) and highly saline (12-22 wt.% NaCl) H2O-CO2 fluids during the magnetite and sulfide stages, respectively (Niiranen et al. 2007). Estimated temperatures for these stages were 450-500°C and 290-370°C respectively and pressure 1.5-3.5 kbars. U-Pb zircon age data on the hanging wall diorite (1864±6 Ma) and the coeval monzonite (1862±3 Ma) give an upper age limit for the mineralisation whereas the U-Pb zircon age of the granite (1760±3 Ma) cross-cutting the ore gives the lower age limit (Hiltunen 1982, Niiranen et al. 2007). Niiranen et al. (2007) proposed, on the basis of the 1797±5 Ma zircon age of the clinopyroxene skarn and a number of ca. 1800 Ma U-Pb ages on titanites in altered host rocks, that the mineralisation took place around 1800 Ma.
The 2.45-2.44 Ga mafic layered intrusions of northern Finland and NW Russia record a global magmatic event with coeval intrusions and dyke swarms occurring in Australia, Canada, India, Antarctica and Scotland. The emplacement of intrusions of this age group is part of a large plume-related rifting event (Alapieti 2005). This event belongs to a global episode of igneous activity in the beginning of the Proterozoic that produced several layered intrusions and mafic dyke swarms on other cratons as well and was, at least in Fennoscandia, related to the initial breakup of an Archaean craton (Alapieti & Lahtinen 2002, Iljina & Hanski 2005).

There are about 20 of these intrusions in Finland; most of them occur along the E-W trending, 300 km long Tornio-Näränkävaara belt in northern Finland (Alapieti et al. 1990; Alapieti & Lahtinen 2002). The intrusions occur either within late Archean tonalitic gneisses or at the contact of the basement gneisses and overlying Early Proterozoic supracrustal sequences. In most cases, the original roof rocks and upper parts of the cumulate units of the intrusions have been obliterated by uplift and subsequent erosion, which took place soon after the intrusion phase. The current roof rocks (< 2.3 Ga volcanic-sedimentary rocks, subvolcanic sills and polymictic conglomerates) were deposited on top of the exposed intrusions. Most of the intrusions consist of more ultramafic basal parts followed by more mafic, gabbroic upper parts. Repetition of ultramafic and mafic layers, i.e. the occurrence of megacyclic units, is a typical feature for the intrusions. Three different types of parental magmas have been recognised based on the composition of chilled margins, cumulates and cogenetic dykes. Two of these have boninitic or siliceous high-magnesian basaltic affinities, with either high- or low-Cr contents. The third magma type was a more evolved tholeiitic basalt (Alapieti et al. 1990, Iljina 1994, Iljina 2005).

The Kemi Cr deposit was found in 1959, during excavation of a fresh-water channel. Full-scale open pit mining started in 1968 and underground operations started in 2003. By the end

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**Figure 10. Cross section of the Kemi intrusion, after Alapieti & Huhtelin 2005.**

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**KEMI Cr DEPOSIT**

Tuomo Törmänen & Tuomo Karinen (GTK)
of 2014 approximately 44 Mt of ore had been mined. The current mineral reserves are 50 Mt @ 26.0 % Cr₂O₃ with an additional 97.8 Mt of mineral resources @ 29.4 % Cr₂O₃ (Outokumpu press release, 2014).

The 2.44 Ga Kemi layered intrusion is 15 km long and 2 km wide. The intrusion is composed of a peridotitic, olivine-chromite cumulate lower part, pyroxenitic (bronzite-augite cumulate) mid-part, and a gabbroic (plagioclase-augite cumulate) upper part (Figure 10). A gabbro-pyroxenite sequence has been encountered in parts of the lower contact with footwall Archean granitoids, probably representing a locally preserved contact series (marginal zone) (Huhtelin 2011). Neoarchean granitoids form the intrusion footwall, and it is capped by younger, 2.15 Ga old mafic volcanics or sub-volcanic sills and by polymictic conglomerate, indicating an erosional upper contact, whereby the original roof rocks and uppermost cumulates were removed. During the Svecokarelian orogeny (1.9-1.8 Ga) the intrusion underwent amphibolite-facies metamorphism, which caused alteration of the original silicate minerals in the lower and upper parts of the intrusion, and tilted the intrusion to its present orientation.

Numerous chromitite horizons occur in the peridotitic lower part of the intrusion, of which the main chromitite layer is the most significant. It occurs approximately 50 – 150 m above the basal contact, but its location has been affected by several strike-slip faults (Alapieti & Huhtelin 2005). It has an average thickness of 40 m, varying from a few meters up to 160 m. The top of the main chromitite is layered, with sharp contacts, whereas the lower contact is more irregular, characterised by chromite disseminations and lumps of chromitite (Huhtelin 2011). The mined part of the main chromite (1.5 km in length) contains several different ore types, based on their gangue mineralogy (e.g., serpentine, chlorite, amphibole), chromite texture and grain size, Cr₂O₃-content, etc. The main products are upgraded lumpy ore and fine concentrate, which are shipped to the Outokumpu Tornio works for ferrochrome and stainless steel production.

KOITELAINEN CHROMITITE LAYERS (LC AND UC)

Tuomo Karinen & Tuomo Törmänen (GTK)

The Koitelainen intrusion belongs to the same age group of 2.45-2.44 Ga old intrusions that include, e.g. the Kemi Cr-deposit, the Mustavaara V-Fe-Ti deposit and the Ahmaavaara PGE-Ni-Cu deposit. The intrusion (Figure 11) is subhorizontal, 26 x 29 km in horizontal extent and ca. 3.2 km in thickness. The footwall rocks of the intrusion are exposed in the anticlinal area in the middle of the intrusion whereas the upper parts of the intrusion are exposed at the margins. The mineralisation at Koitelainen is stratiform. There are two major, sulphide-free, PGE-enriched, chromite reefs possibly extending across the entire intrusion, and a V-rich magnetite gabbro in the intrusion (Mutanen 1997).
abundant thin (< 1 m) pyroxenite layers. The MZ is the thickest unit of the intrusion with an estimated thickness of 1500 m. This zone is mostly composed of gabbro-norites. The 20–40 m thick contact between the MZ and the UZ is composed of heterogeneous anorthosite and chromitite layers of the Upper Chromitite (UC). The main part, i.e. ca. 200 m in thickness, of the UZ is composed of uralite-altered gabbros and anorthosites with the amount of anorthosite decreasing upward. The uppermost part of the UZ is composed of a ca. 100 m thick magnetite gabbro, which is separated from the gabbros below by a unit composed of spotted anorthosite containing ultramafic layers. The magnetite-rich zones are also enriched in vanadium, PGE and, in part, also with copper. (Mutanen 1997)

The Lower Chromitite (LC) layers have been intersected with four diamond drill holes in the Porkausaapa area where the dip of layering is 10° SE. Due to the small amount of drilling it is very difficult to connect individual chromitite layers in the area, but there are probably four to six chromitite layers more than 0.3 m thick, with chromite as the principal phase. The thickest section (>2.0 m) of chromitite in one drill core contains 21.2 wt.% Cr₂O₃. In other diamond drill cores in the Porkausaapa area the Cr₂O₃ content ranges from 10.6 to 32.2 wt.%. In the Porkausaapa area,
the LC has an average Pd+Pt content of 1.4 g/t. (Mutanen 1989). The LC layers have also been intersected in the Rookkijärvi area, located 10 km W of the Porkausaapa area. In this area, the LC layers were intersected with two drill cores in which the individual layers are from few tens of centimeters up to 2.3 meters in thickness. The $\text{Cr}_2\text{O}_3$ content in these layers varies from 8.5 to 27.2 wt.%. The LC in these diamond drill cores also show high Pd+Pt grades, 0.2-1.8 g/t. (Hanski 1998). The locations of the Porkausaapa and Rookkijärvi areas are shown in Figure 11.

The Upper Chromitite (UC) layer was found during studies of the magnetite gabbro and has been intersected by 17 diamond drill holes (Mutanen 1979). According to these cores the UC is a laterally very uniform layer, 150-170 m stratigraphically below the magnetite gabbro. The layer is 1.3 m thick on average and contains 21.0 wt.% $\text{Cr}_2\text{O}_3$, 0.4 wt.% V and 1.1 g/t Pd+Pt. Mutanen (1989) estimated that the mineral resources of the UC are approximately 360 million tonnes of rock with 20.0 wt.% $\text{Cr}_2\text{O}_3$ and 1.1 g/t PGE (Pt+Pd+Rh+Ru+Ir+Os). This estimate is tentative, because it was made by assuming a constant dip of the layering and average compositions for the UC. The estimate was made down to a depth of 500 m from the surface.
AKANVAARA CHROMITITE LAYERS
(LLC, LC, ULC AND UC)
Tuomo Karinen & Tuomo Törmänen (GTK)

The Akanvaara layered intrusion (Figure 13) belongs to the same age group as the 2.45-2.44 Ga old intrusions that include e.g. the Koitelainen, Ahmavaara and Kemi intrusions (see more detailed descriptions of these intrusions in the chapter Kemi Cr deposit). The Akanvaara intrusion hosts a number of layers/zones enriched in various metals including at least 23 chromitite layers, some PGE/Au enriched zones and the V-enriched magnetite gabbro. Of the chromite-rich layers the LC, LLC, ULC and UC (see Figure 14) are the most significant economically (the MC is not considered economically important).

The Akanvaara intrusion is 15 km long and forms a curved north-trending, block-faulted monocline with a surface area of 50-55 km². The floor and roof contacts of the main part of the intrusion have been preserved, although the contacts are faulted in places. At the surface level the intrusion consists of two parts; the main intrusion block including magnetite-rich rocks and the smaller northern "wing", which does not contain similar magnetite-bearing rocks. The main intrusion block can further be divided into three smaller blocks (western, central and eastern), separated by faults (Figure 13). The estimated total stratigraphic thickness of the intrusion is ~3.1 km (Figure 14). Mutanen (1997) divided the layered sequence into three major units: the Lower, Main and Upper Zones, (abbreviated to LZ, MZ and UZ) which have stratigraphic thicknesses of ca. 640 m, 570 m, and 1900 m, respectively. The lower part of the LZ comprises several tens of meters of thick basal, fine-grained gabbro, which is probably the lower chilled margin of intrusion. The intrusion is overlain by a very thick (up to hundreds of meters) layer of granophyres.

The Lowermost Lower Chromitite (LLC) layers occur at the bottom of the LZ, within the heterogeneous subzone of gabbros and pyroxenites adjacent to the lower chilled margin. The Lower Chromitite (LC) layers occur in the homogeneous pyroxenitic rock above the heterogeneous part including LLC layers. In both the LLC and LC, the chromitite layers usually have sharp, faulted lower contacts and gradational upper contacts defined by narrow chromite-pyroxene intervals. The individual layers of LLC and LC generally occur as thin, pinch-and-swell layers or pods showing varying thickness (from decimeters to meters). These layers cannot, therefore, be connected from drill hole to drill hole. The layers of LLC have a maximum thickness of ca. 2 meters. These layers appear also to be disturbed by crosscutting faults, and in some drill holes they are missing altogether. Cr₂O₃ grades in these layers vary generally between 10 to 24 wt.%. The thicknesses of the LC layers locally reach 4-5 m. Cr₂O₃ grades in these layers generally vary from 16 - 34 wt.% (weighted average 24.55 wt.% Cr₂O₃). Both LLC and LC layers are enriched in platinum-group elements (PGE), with typical grades of 200-900 ppb (the weighted average of LLC and LC is ca 600 ppb combined 6 PGE). The Pt/Pd ratio for the LLC is ca 1, and the order of abundance for the PGE and Au is Pd>Ru>Pt>Rh>Ir,Os>Au. The Pt/Pd ratio for the LC is much higher, with an average of 7.3, and the order of abundance for the PGE and Au is Pt>Ru>Os>Rh>Ir>Pd>Au.

The Uppermost Lower Chromitite (ULC) layer occurs close to the lower contact of the MZ, associated with a thin pyroxenite layer at its footwall (Figure 14). The thickness of this individual chromitite layer is usually only a few tens of centimeters: it is associated with highly frac-
tured rock, resulting in high percentage of core loss. The weighted average Cr$_2$O$_3$ content is 22.3 wt.%; the layer has much lower contents of PGE than the LLC and LC (ca. 150 ppb PGE).

The Upper Chromitite (UC) layer, which occurs at the base of the UZ (Figure 14), is economically perhaps the most significant chromite-rich layer in the Akanvaara intrusion. Based on intersections of 30 diamond drill cores it is the most continuous, with thicknesses varying from 0.6 - 1.8 m. The average thickness of the massive part is ca. 1 m with an average grade of 22.79 wt.% Cr$_2$O$_3$. The UC is also enriched in vanadium (0.3-0.52 wt.%) which resides in chromite, and in PGEs (0.915 g/t). Mutanen (1998) estimated the mineral resources of the UC but his estimate must be considered as tentative, since it was made on the basis of a limited number of diamond drill cores. Thirty-eight holes were drilled (total 4994.1 m) targeting the UC layer along its known 6-7 km strike, and assuming the layering dipped at 30° down to 300 m below the surface. There is thus no actual data for the assumed deeper extensions of the UC. According to the estimate, the UC includes 18.1 million metric tons of rock containing 22.8 wt.% Cr$_2$O$_3$, 0.4 wt.% V and 0.912 g/t combined Pd+Pt+Ru+Os+Rh+Ir.
Figure 14. Geological map of the Akanvaaara intrusion with GTK drill holes marked (after Mutanen 1997).

The Portimo Complex is composed of four principal structural units: Konttijärvi, Suhanko, Narkaus intrusion (Lihalampi, Kilvenjärvi, Nuturalampi, Kuohunki and Siika-Kämä blocks) and Portimo Dykes (Figure 15). Each intrusion contains a marginal series and an overlaying layered series. The marginal series generally varies from ten to several tens of meters in thickness. In the Narkaus intrusion the marginal series is mainly composed of pyroxenite while the upper half of the Suhanko and Konttijärvi marginal series is composed of olivine cumulates. The layered series of the Narkaus intrusion contain up to three olivine-rich ultramafic cumulate layers (MCU I - MCU III), whereas the Suhanko and Konttijärvi intrusions contain only one, in the bottom of the layered series. The principal mineralisation types in the Portimo Complex are: 1) the PGE-enriched Cu-Ni-Fe sulphide disseminations in the marginal series of the Suhanko and Konttijärvi intrusions, 2) the massive pyrrhotite deposits located close to the basal contact of the Suhanko intrusion, 3) the Rytikangas PGE Reef in the layered series of the Suhanko intrusion, 4) the Siika-Kämä PGE Reef in the Narkaus layered series, and 5) the offset Cu-PGE mineralisation below the Narkaus intrusion.

The Konttijärvi Intrusion is 1.2 km long and 0.5 km wide at the surface section and dips 35-60° N: its footwall and hanging wall rocks consist of Archaean migmatisitic mica gneiss (Figure 15). The intrusion has been divided into two principal units: the marginal series and the layered series (Figure 16). The marginal series is ca. 40 m thick and the layered series ca. 120 m thick. The lower part of the marginal series is composed of pyroxenitic cumulates and the upper part of peridotitic olivine cumulates. The layered series is composed mainly of gabbronoritic cumulates with a ca. 25 m thick orthopyroxenite layer in the middle part of it.

The disseminated PGE-enriched, marginal series base-metal sulphide deposit of the Konttijärvi intrusion is generally 10–30 m thick (Figures 16 and 17, Table 1). It is, however, erratic and commonly extends from the lower peridotitic layer downwards for some 30 m into the basement gneiss (peridotite → pyroxenite → gabbro → basement gneiss). The PGE contents vary from only weakly anomalous values to 2 ppm but some samples contain > 10 ppm PGE (Figures 16 and 17).

The Suhanko intrusion is, at the surface, about 16 km long and 1 km wide, dipping 5-70° to the NE, SW or W, depending on the part of the intrusion (Figure 15): its footwall rocks consist of Archaean diorite and felsic volcanic rocks. The hanging wall rocks, which are younger than the intrusion, include tonalitic migmatite, mafic vol-
Figure 15. Generalised geological map of the Portimo Complex. Modified after Iljina 1994.

Figure 16. Generalised geological cross-section of the Konttijärvi layered intrusion. Modified after Iljina 1994.
Figure 17. Stratigraphic sequence of the Kontti-
jarvi marginal series showing variations in
bulk Pd, Cu, Ni and S in diamond drill hole KO-3.
Modified after Alapieti et al. 1989.

Figure 18. Stratigraphic sequence of the
Ahmavaara marginal series showing varia-
tions in bulk Pd, Cu, Ni and S in diamond drill
canic rocks, conglomerates and quartzites. The intrusion has been divided into two principal units: the marginal series and the layered series. The marginal series is 10-60 m thick (see Figure 18) and the layered series about 500 m thick. The lower and middle parts of the marginal series are composed of pyroxenitic and gabbronoritic cumulates and the upper part of peridotitic olivine cumulates. The layered series is composed mainly of gabbronoritic and anorthositic cumulates.

Massive sulphide mineralisations occur in the Suhanko intrusion as dykes and plate-like bodies conformable to the layering. The thickness of the massive sulphide mineralisations varies from 0.2 - 20 m: they are located within the interval of 30 m below the basal contact of the intrusion and 20 m above it (see Figure 18, Table 1). The Ahmavaara sulphide paragenesis is composed of pyrrhotite, chalcopyrite and pentlandite. Normally the massive pyrrhotite deposits of the Suhanko Intrusion show low PGE values (max. a few ppm Pt + Pd). The PGE concentrations are generally much higher in the Ahmavaara deposit (up to 20 ppm).

The Narkaus intrusion (Lihalampi, Kilvenjärvi, Nutturalampi, Kuohunki and Siika-Kämä blocks) is about 22 km in total length, up to 2.5 km in width and its footwall rocks consist of Archaean granitoids. The hanging wall rocks, which are younger than the intrusion, include
mafic volcanic rocks, conglomerates, quartzites, mica schists and Archaean granitoids as in the case of Siika-Kämä block. The intrusion has been divided into two principal units: the marginal series and the layered series. The marginal series of the Kilvenjärvi block (Figure 19), about 10 m thick, is composed of gabbronoritic and bronzititic cumulates and partly melted floor-rock xenoliths. The layered series, about 860 m thick, is composed of alternating sequences of ultramafic, gabbronoritic, gabbroic and anorthositic cumulates and it has been divided into megacyclic units I - III, starting from the base and abbreviated as MCU I – MCU III. The chemical composition of megacyclic units I – II is boninitic (Cr-rich) and of megacyclic unit III tholeiitic (Cr-poor) (see Figure 20). These variations are interpreted as being attributable to repeated influxes of new magma into the Narkaus chamber during solidification.

The Siika-Kämä PGE Reef (Table 1) of the Narkaus intrusion is located at the contact zone between the gabbronorite of MCU II and the ultramafic cumulates of MCU III, most commonly at the base of MCU III, but in places it may lie somewhat below that or in the middle of the olivine cumulate layer of MCU III (Figure 20). The Siika-Kämä Reef has been identified in the Kilvenjärvi, Nurturalampi, Kuohunki and Siika-Kämä intrusion blocks (Figure 15). Chlorite-amphibole schist is the most common host rock for the Siika-Kämä Reef. Normally the reef contains weak base metal sulphide disseminations, but locally also chromite seams or chromite disseminations. The thickness of the reef varies from less than one meter to several meters: many drill holes penetrate a number of mineralised layers. The reef PGE concentration varies from several hundred ppb to tens of ppm.

![Figure 20. Whole-rock chemical variation in the Siika-Kämä PGE Reef. Modified after Huhtelin et al. 1989.](image)

### Table 1. PGE±Cu-Ni deposits with a resource estimate in the Portimo Complex area.

<table>
<thead>
<tr>
<th>Occurrence</th>
<th>Tonnage (Mt)</th>
<th>Pd g/t</th>
<th>Pt g/t</th>
<th>Cu %</th>
<th>Ni %</th>
<th>Au g/t</th>
<th>Deposit subtype</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Konttijärvi</td>
<td>75.3</td>
<td>0.95</td>
<td>0.27</td>
<td>0.1</td>
<td>0.05</td>
<td>0.07</td>
<td>Contact</td>
<td>Puritch et al. (2007)</td>
</tr>
<tr>
<td>Ahmavaara</td>
<td>187.8</td>
<td>0.82</td>
<td>0.17</td>
<td>0.17</td>
<td>0.09</td>
<td>0.9</td>
<td>Contact</td>
<td>Puritch et al. (2007)</td>
</tr>
<tr>
<td>Siika-Kämä</td>
<td>43.1</td>
<td>2.7</td>
<td>0.72</td>
<td>0.11</td>
<td>0.08</td>
<td>0.08</td>
<td>Reef</td>
<td>Gold Fields (2003)</td>
</tr>
</tbody>
</table>
The Mustavaara V-Fe-Ti oxide deposit is hosted by the Koillismaa Intrusion, which is a part of the ca. 2440 Ma Tornio-Näränkävaara Intrusion belt (see more detailed description of these intrusions in the chapter Kemi Cr deposit). The ore in the Mustavaara deposit is defined by the oxide-rich lower part of the magnetite gabbro layer within the upper part of the Koillismaa Intrusion (“magnetite-rich cumulates” in Figure 21, see also Figures 22 and 23). This block is one of the largest blocks of the intrusion, and it has been modelled gravimetrically to extend to a depth of ~ 2000 m. The block is 20 km long and 4 km thick with an exposed area of ~ 80 km². The magnetite gabbro layer is uniform and can be traced in almost every block of the Koillismaa Intrusion. The layer is, however, thicker in the Porttivaara block area than in the other intrusion blocks and in the Mustavaara deposit area it reaches a thickness of 240 m (Juopperi 1977, Ruotsalainen 1977, Piirainen et al. 1978, Alapieti 1982, Iljina & Hanski 2005, Karinen et al. 2015).

According to Juopperi (1977) and Karinen et al. (2015) the Mustavaara deposit comprises a ca. 80 m thick succession of ilmenomagnetite-rich sub-layers in the lower part of the magnetite gabbro layer. The grades of the ore in these layers correlate with the amount of the ilmenomagnetite, an oxide phase which originally crystallised as titaniferous magnetite, but which later, in a subsolidus stage reacted to form composite grains of fine ilmenite lamellae in a V-bearing magnetite host (Table 2, Figures 22 and 23). The dip of the ore layer in the area of the deposit is 40° N, but steepens to 60° east of the old open pit, where the layer also becomes narrower, being only 20 m thick. Depending on the amount of ilmenomagnetite, the deposit is divided into distinct sub-layers comprising, from the bottom upwards: the Lower Ore Layer (LOL; 5 m thick; 20–35 wt.% ilmenomagnetite), the Middle Ore Layer (MOL; 15–50 m thick; 10–15 wt.% ilmenomagnetite), the Upper Ore Layer (UOL; 10–40 m thick; 15–25 wt.% ilmenomagnetite) and the Disseminated Ore Layer (DOL; <10 wt.% ilmenomagnetite).

The history of mining in the Mustavaara deposit dates back to the mid 1950s when a local man observed compass interferences in the area adjacent to the deposit. This encouraged him to send samples from the area to different mining companies. The samples led the Otanmäki Company to launch an exploration project in the area. The decision was also facilitated by the aeromagnetic anomalies that were detected in the area during...
Figure 21. Generalised geological map of the Koillima-Närünkävaara Layered Complex. Modified from Karinen (2010).

Figure 22. Plan view of the Mustavaara open pit area showing the ore layers projected to surface.
the high altitude survey program carried out by the Geological Survey of Finland (GTK) (1951–1972). The measurements showed that the samples represent a coherent high anomaly layer in the Mustavaara area. During the following years, the Otanmäki Company explored the area. In 1967 this work led to the discovery of a V-bearing magnetite gabbro layer in the upper part of the Porttivaara block of the Koillismaa Intrusion. (Isokangas 1957, Juopperi 1977, Markkula 1980).

The decision to open a mine in the Mustavaara area was made by the Rautaruukki Steel Company in 1971 (the Otanmäki Company was merged with Rautaruukki in 1968). Open-pit mining began in 1976, but was terminated in 1985 due to the low price of vanadium. The ore reserves of the Mustavaara Mine were 38 Mt at 16.8 wt.% ilmenomagnetite concentrate, with a cut-off value of 11.9 wt.% (Paarma 1971). The reserves were estimated to 100 m below the surface. During its operational life, the mine produced 13.45 million tons of ore and 1.97 million tons of ilmenomagnetite-rich concentrate averaging 0.91 wt.% V. During the operation of the mine, one significant challenge to overcome was the optimisation of the production line to gain the maximum V-grade with minimum amount of ilmenite lamellae in the magnetic concentrate. The annual production was approximately 240,000 tons of concentrate and 2,500–3,000 tons of V₂O₅ (1400–1700 t V), accounting for 6–9 % of the global production of vanadium at that time (Juopperi 1977, Puustinen 2003).

The mine area has recently been re-evaluated for additional ore potential by Mustavaara Mine Ltd. A drilling programme by the company has outlined a down-dip continuation of the magnetite gabbro. A new ore reserve estimate has been calculated to a depth of 250 m down from the topographic surface. The current reserves are 99 million tons of ore grading 14.0 wt.% ilmenomagnetite with a vanadium content of 0.91 wt.%. The reserves include 64 million tons in proven and 35 million tons in the probable class. These estimates were calculated using an ilmenomagnetite cut-off value of 8.0 wt.% (Mustavaaran Kaivos Oy, 2013).

Figure 23. Vertical cross section of the ore of the Mustavaara Deposit along profile 10200. See location of the profile in Fig. 22.
THE KEVITSA AND SAKATTI Ni-Cu-PGE DEPOSITS
Tuomo Törmänen & Tuomo Karinen (GTK)

The Kevitsa and Sakatti Ni-Cu-PGE deposits are located within the Central Lapland greenstone belt (CLGB) in northern Finland. The CLGB consists of volcano-sedimentary sequences, spanning a time interval from the 2.44 Ga Salla Group felsic-intermediate metavolcanic rocks to ca. 1.88 Ga molasse-type metasediments of Lainio and Kumpu Groups (Hanski & Huhma 2005). Mafic magmatism within the CLGB is recorded by a number of intrusive phases. The oldest mafic intrusive phase is represented by the 2.44 Ga mafic layered intrusions of Koitelainen and Akanvaara (Figure 24). At 2.20 Ga a suite of differentiated mafic sills intruded into the southern part of the CLGB. 2.15 Ga gabbroic sills and dykes occur in the central and south-eastern part of the schist belt. The last major mafic intrusive phase is represented by the 2.05-2.0 Ga Kevitsa-type layered intrusion and sills, as well as minor gabbroic intrusions and dykes (Rastas et al. 2001; Räsänen & Huhma 2001; Hanski & Huhma 2005). The 2.44 Ga layered intrusions host a number of meter-scale chromitite (+PGE) horizons as well as V-enriched magnetite gabbros. The only known magmatic Ni-Cu-PGE deposits are hosted by the 2.05 Ga gabbroic sills and dykes (Rastas et al. 2001; Räsänen & Huhma 2001; Hanski & Huhma 2005). The 2.44 Ga layered intrusions consist of ultramafic, gabbroic, and granophyric parts with locally developed mafic layering. The ultramafic part, which hosts the mineralisation, occurs at depth and in the NE-part of the intrusion. It is mostly composed of olivine websterite, olivine clinopyroxenite and minor plagioclase-bearing (olivine) websterite, and a pyroxenite zone between the ultramafic and gabbroic parts of the intrusion (Santaguida et al. 2015). The cumulate rocks of the ultramafic part are variably altered to amphibole-chlorite-serpentinite-bearing assemblages. Gabbroic rocks (gabbro, olivine gabbro, and gabbronorite) form the upper (SW) part of the intrusion. The granophyre zone, according to Mutanen (1997), forms the southern part of the intrusion, but this has not been encountered in the FQM drill holes. The marginal zone of the intrusion is made up of pyroxenite and gabbro with a gradational contact to olivine websterite. A km-sized ultramafic body occurs in the central part of the intrusion. Mutanen (1997) describes this as a serpentinised dunite xenolith, unrelated to the intrusion. Numerous ultramafic and sedimentary xenoliths occur throughout the intrusion and especially in the mineralised part of the intrusion (Mutanen 1997; Santaguida et al. 2015). The immediate country rocks of the intrusion consist of mafic volcanic rocks, phyllite with graphite- and sulphide-rich interlayers, arkosic quartzite, and carbonate-bearing schists (Mutanen 1997; Santaguida et al. 2015).

Kevitsa deposit
The Kevitsa Ni-Cu-PGE deposit is located in northern Finland, 170 km N of the Arctic Circle, within the Paleoproterozoic Central Lapland greenstone belt (CLGB). The deposit was originally discovered by the Geological Survey of Finland in 1987 (Mutanen 1997), with continued exploration until mid 1990s. The deposit was subsequently explored by Outokumpu Ltd, Scandinavian Minerals, and finally by First Quantum Minerals Ltd. (FQM), which started production in 2012. The current (2012) mineral resource is 240 Mt at 0.3 % Ni, 0.41 % Cu, 0.21ppm Pt, 0.15ppm, Pd, and 0.11ppm Au (Santaguida et al. 2015).

The disseminated Ni-Cu-PGE mineralisation is hosted by the 2.06 Ga (Mutanen & Huhma 2001) layered or composite Kevitsa intrusion. The intrusion consists of ultramafic, gabbroic, and granophyric parts with locally developed mafic layering. The ultramafic part, which hosts the mineralisation, occurs at depth and in the NE-part of the intrusion. It is mostly composed of olivine websterite, olivine clinopyroxenite and minor plagioclase-bearing (olivine) websterite, and a pyroxenite zone between the ultramafic and gabbroic parts of the intrusion (Santaguida et al. 2015). The cumulate rocks of the ultramafic part are variably altered to amphibole-chlorite-serpentinite-bearing assemblages. Gabbroic rocks (gabbro, olivine gabbro, and gabbronorite) form the upper (SW) part of the intrusion. The granophyre zone, according to Mutanen (1997), forms the southern part of the intrusion, but this has not been encountered in the FQM drill holes. The marginal zone of the intrusion is made up of pyroxenite and gabbro with a gradational contact to olivine websterite. A km-sized ultramafic body occurs in the central part of the intrusion. Mutanen (1997) describes this as a serpentinised dunite xenolith, unrelated to the intrusion. Numerous ultramafic and sedimentary xenoliths occur throughout the intrusion and especially in the mineralised part of the intrusion (Mutanen 1997; Santaguida et al. 2015). The immediate country rocks of the intrusion consist of mafic volcanic rocks, phyllite with graphite- and sulphide-rich interlayers, arkosic quartzite, and carbonate-bearing schists (Mutanen 1997; Santaguida et al. 2015).
The irregularly shaped disseminated Ni-Cu-PGE deposit is concentrated in the central (ultramafic) part of the intrusion. It extends from the surface to a depth of approximately 400 metres. Mutanen (1997) recognised four different types of sulphide mineralisation: the regular ore, which makes up the bulk of the Ni-Cu-PGE mineralisation, Ni-PGE ore, transitional ore, and false ore. In the current FQM classification the Ni-PGE and transitional ore types are classified as Ni-PGE ore (Santaguida et al. 2015). The regular ore is composed of interstitial, disseminated sulphides, mainly pyrrhotite, pentlandite and chalcopyrite, with a Ni tenor from 4-7 % and <1 ppm PGE. The regular ore has Ni/Cu ratios of ca. 0.5 and Pt/Pd ratios >1: Ni-Cu concentrations vary, however, at local and deposit scale with more Ni-rich zones occurring at depth and in the southern part of the deposit, and Cu-rich zones in the central parts of the deposit. Ni-PGE ore, which constitutes approximately 5 % of the resource, occurs as discordant zones within the regular ore. It has a higher Ni tenor (>10 %) and higher PGE contents. It is also typified by the presence of heazlewoodite, millerite, and pyrite as additional sulphide minerals. False ore generally contains more sulphides, varying from heavy disseminations to net-textured to semi-massive pyrrhotite-rich ore with only minor pentlandite and chalcopyrite resulting in a low Ni tenor of 2-3 % (Mutanen 1997; Santaguida et al. 2015).

**Sakatti deposit**

The Sakatti Ni-Cu-PGE deposit is located in northern Finland, 150 km N of the Arctic Circle, within the Paleoproterozoic Central Lapland greenstone belt (CLGB). Regional exploration by Anglo American led to the discovery of the deposit in 2009 (Coppard 2011).

Currently, three separate mineralised ultramafic bodies are known to occur in the area: The main body, NE-body and SW-body. The main body subcrops in a ca. 250x500 m area. It forms an irregular, 200-400 m wide, NW-plunging, tubular intrusion with a known depth extent of approximately 1200 m (Brownscombe et al. 2015). The intrusion consists mostly of serpentiniﬁed peridotite (olivine cumulate with interstitial clino- and orthopyroxene) and subordinate dunite (olivine adcumulate) at depth. On a regional scale, the ultramafic bodies are surrounded by quartzites of the Sodankylä Group (≥2.2 Ga) and pelitic metasediments of the Savukoski Group (ca. 2.06 Ga Matarakoski Formation). The immediate wall rocks to the main intrusion, however, are not easily classified into either of the aforementioned groups. The most prominent wall rock is a fine grained, aphanitic unit, which is composed of small olivine and plagioclase phenocrysts in a fine grained plagioclase-pyroxene-olivine groundmass. Close to the contact with the intrusion it contains mm- to cm-scale fingers or veins of intrusion peridotite. Based on mineralogical and textural features, it is thought to be volcanic in origin, compositionally resembling komatiite-picrite. In the SW part of the deposit, wall rocks include a heterogeneous package containing mafic volcanic rock, scapolite-biotite schist and gabbroic sills intruding the mafic volcanite and scapolite-biotite schist (so called mafic suite). Upwards in the stratigraphy, the aphanitic unit and the mafic suite are followed by a 100-300 m thick hematite-carbonate-albite-talc altered polymictic breccia unit. Uppermost in the wall rock sequence there is a 600 m thick phyllitic unit (Brownscombe et al. 2015).

The sulphide mineralisation consists of disseminated sulphides, sulphide veins, and semi-massive to massive sulphides.

**Disseminated sulphides** extend from the subcropping SE-part of the intrusion, through the central part of the intrusion to the deeper parts, although the whole of the intrusion is not mineralised. The disseminated sulphides are mainly composed of chalcopyrite with minor pyrrhotite, pentlandite, and pyrite. The disseminated sulphides tend to be more Cu-rich in the shallower parts and Cu-poor in the deeper parts (Brownscombe et al. 2015). Typical grades along thick intersections are 0.51 % Cu, 0.23 % Ni, 0.44 ppm Pt, 0.22 ppm Pd, and 0.13 ppm Au (62.70 m in DDH 08MOS8007) (Coppard et al. 2013).

**Massive to semi-massive sulphide veins** occur predominantly in the shallow and central parts of the intrusion (together with disseminated sulphides). They are Cu-rich, Au-enriched (compared to disseminated and massive sulphides) and typically 5-20 cm thick (Brownscombe et al. 2015; Coppard et al. 2013).

**Massive sulphides** occur as stacked lenses in the...
central part of the intrusion, extending up to 150 m into the aphanitic sidewall. The lenses are up to 25 m thick in the central part, thinning down to 0.5 m towards NW and SE. A 26.5 m interval of massive sulphides (DDH 11MOS8049) contained 3.69 % Cu, 4.16 % Ni, 0.18 % Co, 1.10 ppm Pt, 1.27 ppm Pd, and 0.24 ppm Au (Coppard et al. 2013). Disseminated and vein sulphides have low Ni/Cu ratios and high Pt/Pd ratios of <1 and 2, respectively. The massive sulphides show highly variable Ni/Cu ratios, with high ratios (Ni/Cu>10) in the deeper part of the intrusion and more Cu-rich in the upper parts of the intrusion (Ni/Cu<0.1). Platinum dominates over palladium in the disseminated and vein-style mineralisations with Pt/Pd ratios of ca. 2. In the massive sulphides, Cu-rich parts are Pt-rich, whereas Ni-rich parts are more Pd-rich (Brownscombe et al. 2015). No grade-tonnage estimates have been published for the Sakatti deposit(s) but based on published data it is potentially a world-class deposit, if not giant or even super giant.

Figure 24. Location of 2.44 Ga layered intrusions (Koitelainen, Akanvaara) and the Kevitsa and Sakatti Ni-Cu-PGE deposits within the Central Lapland Greenstone Belt (CLGB). 2.2 Ga layered sills occur in the Kevitsa-Sakatti area and at the contact between CLGB and the Central Lapland Granitoid Complex (CLGC).
The Otanmäki Fe-Ti-V oxide deposits are hosted by a mafic to anorthositic intrusion complex (2065 ± 4 Ma) which is situated along with alkaline granitoids (2050 Ma) at the boundary between the Archaean Pudasjärvi and Isalmi blocks, immediately W of the Palaeoproterozoic Kainuu schist belt (Talvitie & Paarma 1980, Konttinen et al. 2013a). In addition to ferrous metals, the area is a potential source for REE, Zr and Nb in gneissic alkaline granitoids. The belt of intrusive alkaline granitoids extends a few tens of kilometers to the east (Figure 25).

The Otanmäki ferrous-metal deposits are orthonmagmatic magnetite-ilmenite ores that have undergone a complex deformation and meta-morphism. For most parts, the Otanmäki and Vuorokas intrusions are tectonised and show strong metamorphic fabrics (foliation). The gabros are layered, and well developed magmatic differentiation is only seen in unmetamorphosed parts of the deposits. The oxide ore is located in the heterogeneous gabbro-anorthosite unit which is the stratigraphical Upper part of the intrusion. The Lower part is mainly seen in Vuorokas: it consists predominantly of gabbronorites, olivine gabbronorites and bronzite cumulates (Nykänen 1995).

The Otanmäki mine operated from 1953–1985 (Vuorokas Mine 1979–1985). In total, 30 Mt of ore grading of 32-34 % Fe, 5.5-7.6 % Ti and 0.26 % V was mined (Puustinen 2003). The processing plant in Otanmäki produced magnetite (7.6 Mt), ilmenite (3.8 Mt), sulphur (0.2 Mt) concentrates and vanadium pentoxide (55 545 t) from the vanadium plant. Vanadium pentoxide (V₂O₅) was the main product for most of the
mine life (Illi et al. 1985). The mine consists of three separate shafts (Otanmäki, Suomalmi, and Vuorokas), and 125 kilometers of tunnels. The deepest level in Otanmäki reaches down to the +675 level (485 m above sea level).

The heterogeneous nature of the Otanmäki deposit is clearly seen in its many unequally sized and irregularly shaped ore lenses (Figure 28) which contain abundant gangue inclusions. The lenses are subvertical, 2–200 m long and 3–50 m thick, strike E-W and dip at 70°–90° N. The mineralised zone (Figure 26 a) is roughly 3 km long and 0.5 km wide, and it forms a semicircle-shaped syncline structure (Figures 26 and 27) at its eastern end. The syncline structure has a plunge of ca. 40°–60° to SW-W and the ore zone is concordant with the fold axes. Foliation cuts the primary magmatic layering normally in 1–20° angles. The magmatic layering is mainly controlling the ore but due to the deformation axis of individual lenses tends to face the line-
The ore zone is open at depth but is known to extend to >800 m (Lindholm & Anttonen 1980).

The average mineralogy of the Otanmäki (Figure 29) high-grade ore consists of magnetite (35–40 wt. %), ilmenite (25–30 wt. %) and sulphides (1–2 wt. %). The main gangue minerals are chlorite, hornblende and plagioclase. Magnetite and ilmenite occur mainly as granoblastic textured, separate 0.2–0.8 mm grains (Kerkkonen 1979, Pääkkönen 1956). In parts, ilmenomagnetite is predominant, containing ilmenite and spinel as exsolved lamellae and inclusions in magnetite (Figure 29) (Kerkkonen 1979).

The vanadium content in the magnetite is 0.80–1.18 % $V_2O_3$ (~ 0.62 wt. % V) and varies slightly between the ore bodies (Kerkkonen 1979, Pääkkönen 1956).

Another style of metallic mineralization within...
the area is defined by rare metal and REE prospects, mainly associated with alkaline gneisses. The Katajakangas Nb-REE prospect (0.46 Mt at 2.4 % RE_2O_3, 0.31 % Y_2O_3, and 0.76 % NbO) is a narrow mineralised zone hosted by sheared quartz-feldspar gneisses with riebeckite and alkaline pyroxene (Hugg 1985a, Puumalainen 1986, Al-Ani et al. 2010). The main ore minerals are fergusonite, allanite and columbite (Hugg 1985a). The fergusonite is characterized by a high content of radioactive elements such as Th and U. Katajakangas has clearly a metasomatic origin (Al-Ani et al. 2010, Sarapää et al. 2013). The Kontioaho occurrence, another REE mineralisation in the Otanmäki area, contains 1.62 Mt at 3.3 % ZrO_2, 0.66 % Ln_2O_3, 0.18 % Y_2O_3, and 0.14 % Nb_2O_5. Kontioaho is situated 1.3 km NNE of Katajakangas. The surface projection of the mineralized zone is approximately 400 x 600 m and it forms a 12 m thick stratiform vein-like body at the basal contact of quartz-feldspar schist (Hugg 1985b). The main ore mineral assemblage includes fergusonite, allanite, and xenotime (Hugg 1985b).

In general, the Katajakangas and Kontioaho mineralisations contain fergusonite (Nb, Y, HREE), allanite (LREE), columbite (Nb), and xenotime (Y). In addition, titanite in Kontioaho contains high amounts of yttrium (Äikäs 1990).
LUMIKANGAS AND PERÄMAA DEPOSITS

Niilo Kärkkäinen (GTK)

Lumikangas Fe-Ti-P gabbro

The Lumikangas gabbro is the northernmost intrusion of the Kauhajoki Fe-Ti-P gabbro field (Peräkorpi Ti metallogenic area in Eilu et al. 2012) situated between synkinematic (ca. 1.88 Ga) and postkinematic (ca. 1.87 Ga) granitoids in the western part of Central Finland Granitoid Complex (Figure 30). The Geological Survey of Finland (GTK) found and studied the Lumikangas Fe-Ti-P gabbro using geophysical surveys and drilling during the period 2002-2004. The reasons for initiating these studies were 5 km long aeromagnetic and regional gravity anomalies in an unexposed area with a 30-70 m thick Quaternary soil cover (Sarapää et al. 2005). The Fe-Ti-P Kauhajoki gabbros and the Lauhanvuori granite probably belong to a post-orogenic bimodal magmatic suite (Kärkkäinen and Appelqvist 1999, Rämö et al. 2001, Peltonen 2005b).

The Lumikangas intrusion is layered with a basal part composed of dark medium-grained gabbro, hornblende gabbro and olivine gabbro, and an upper part of medium to coarse-grained leucogabbro and monzogabbro. The layers have a gentle dip, 30° to the east. Ilmenite occurs as separate grains, in anhedral to subhedral granular aggregates with magnetite and as exsolution lamellas in magnetite. (Chernet et al. 2004). The vanadium content of the magnetite varies from 0.3 - 1.6 % V. The Lumikangas gabbro is characterised by rather high K2O and low Cr contents, a high alkali feldspar component in normative feldspar, and coeval crystallisation of apatite, Fe-Ti oxides and mafic silicates (Sarapää et al. 2005).

The Lumikangas gabbro contains disseminated ilmenite, apatite and Ti-magnetite, in total averaging 18–22 wt.%, distributed with slightly variable amounts throughout the intrusion (Sarapää et al. 2005). Magnetic and gravity surveys indicate that the Lumikangas deposit extends to a depth of 300 - 500 m: the total length of the anomalies is 5 km. The possible resources include 230 million tons of oxide gabbro containing an average of 18 % ore minerals: 8.4 % (max. 21 %) ilmenite, 4.3 % (max. 17 %) magnetite and 5.0 % (max. 13 %) apatite. The estimate is based on geophysical interpretation and two drilling sections (totalling 1308 m), according to which the Fe-Ti oxide rich block (with specific gravity 3.2 t/m³) is 1200 m long, 300 m wide and 200 m deep.

Perämaa Fe-Ti-P gabbro

The Perämaa gabbro is the major potential source of iron, phosphorus and titanium in the Kauhajoki Fe-Ti-P gabbro field (also called the Peräkorpi Ti metallogenic area in Eilu et al. 2012) (Figure 30). It was found by drilling an aeromagnetic anomaly within the western part of the Central Finland Granitoid Complex. Several Fe-, Ti- and P-rich mafic intrusions rim the postorogenic Honkajoki pluton which consists of potassium-rich granites, forming a bimodal magmatic suite with the Fe-Ti-P gabbros (Rämö et al. 2001, Peltonen 2005b).

Exploration in the Perämaa area has been limited. During the discovery drilling (totalling 411 m) by the Geological Survey of Finland (GTK) the major interest was in iron and titanium. In the early 1980s the possibility of mining phosphorus ore was studied by Kemira Oy in cooperation with GTK (ground geophysics, 2845 m drilling, beneficiation tests) (Pakarinen 1984, Ervamaa 1986).

The low-grade Perämaa Fe-Ti-P deposit is composed of disseminated and thin semi-massive layers of ilmenite, titaniferous magnetite and...
apatite in rhythmically layered peridotites, olivine gabbro-rorites, gabbro-rorites and gabbros, in which the layering is almost vertical (Pakarinen 1984, Rämö 1986). Ilmenite occurs both as separate grains and as lamellae in magnetite (ilmenomagnetite). In the Perämaa intrusion, there are three Fe-Ti-P-mineralised ultramafic-mafic blocks, which are estimated to contain 200-300 million tons mineralised rock containing about 2.5 % P₂O₅, 5 % TiO₂ and 10 % magnetite (with 0.5 % V) down to depth 60-120 m (Ervamaa 1986, Pakarinen 1984). The depth of the intrusion is interpreted, on the basis of gravity and magnetic surveys, to be more than 400 m. The Perämaa Fe-Ti-P gabbro can thus be rated as a potential future resource for apatite, ilmenite, iron and vanadium.
The Archaean Siilinjärvi carbonatite-glimmerite complex is located in the Archaean Karelia province near the boundary of the Palaeoproterozoic Svecofennian domain. It is the oldest carbonatite to be mined for phosphate fertilizer as well as being one of the oldest carbonatites on Earth. The complex is a steeply dipping lenticular body roughly 16 km long with a maximum width of 1.5 km and a surface area of 14.7 km². The carbonatite-glimmerite occurs as tabular bodies of glimmerite and carbonatite intruded into Archaean basement gneiss, which shows a fenitised halo around the bodies (Puustinen 1971). Two of the carbonatite-glimmerite bodies are currently being mined for apatite, the main pit Särkijärvi in the south (see Figures 31 and 32) and the satellite pit Saarinen, in the north. The abundant zircon found in all the rock types in the complex has allowed several U-Pb age determinations to be made for the Siilinjärvi carbonatite-glimmerite, giving reliable ca. 2.61 Ga age results (e.g. Kouvo unpublished; Rukhlov & Bell 2010, Tichomirowa et al. 2013, Zozulya et al. 2007).

The carbonatite and glimmerite are intimately mixed, varying from nearly pure glimmerites (tetraferriphlogopitites; Puustinen 1974) to nearly pure carbonatites. Well-developed sub-vertical to vertical lamination can be observed throughout the main lithologies. Apatite is observed in all rock varieties. Although the complex is not distinctly zoned, the volume of carbonatite is greatest near the center of the intrusion. Glimmerites near the outer edges of the body can be nearly carbonate-free, but still contain ore-grade amounts of apatite (Table 3). Two types of carbonatite can be observed; an apatite-bearing type (with a 4:1 calcite:dolomite ratio) and a younger, transecting apatite-poor type (with a 4:1 calcite:dolomite ratio; O’Brien et al. 2015). The compositions of both carbonatites are relatively pure calcio-carbonatite. Light green to grayish apatite forms rounded grains or euhedral hexagonal rods, up to several centimetres long. All the apatite found at Siilinjärvi is fluorapatite, containing 0.75 - 4 wt. % F (Hornig-Kjarsgaard 1998, O’Brien et al. 2015).
Table 3. Siilinjärvi ore zone rocks, modal mineralogy, and major elements chemistry after O’Brien et al. (2015).

<table>
<thead>
<tr>
<th></th>
<th>Ore¹</th>
<th>Glimmerite</th>
<th>Carbonatite, apatite-bearing</th>
<th>Carbonatite, apatite-poor</th>
</tr>
</thead>
<tbody>
<tr>
<td>Micas²</td>
<td>65.0</td>
<td>81.5</td>
<td>14.7</td>
<td>1.7</td>
</tr>
<tr>
<td>Amphibole</td>
<td>5.0</td>
<td>4.5</td>
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<td>0.2</td>
</tr>
<tr>
<td>Calcite</td>
<td>15.0</td>
<td>1.6</td>
<td>1.7</td>
<td>0.2</td>
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<tr>
<td>Dolomite</td>
<td>4.0</td>
<td>0.9</td>
<td>13.4</td>
<td>10.6</td>
</tr>
<tr>
<td>Apatite</td>
<td>10.0</td>
<td>10.4</td>
<td>9.9</td>
<td>0.8</td>
</tr>
<tr>
<td>Accessories</td>
<td>1.0</td>
<td>0.7</td>
<td>0.1</td>
<td>0.4</td>
</tr>
<tr>
<td>SiO₂ (wt. %)</td>
<td>30.2</td>
<td>37.5</td>
<td>7.8</td>
<td>1.3</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.3</td>
<td>0.5</td>
<td>0.1</td>
<td>&lt;0.1</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>7.0</td>
<td>8.8</td>
<td>1.8</td>
<td>0.2</td>
</tr>
<tr>
<td>Fe₂O₃t</td>
<td>7.1</td>
<td>8.3</td>
<td>3.0</td>
<td>1.6</td>
</tr>
<tr>
<td>MnO</td>
<td>0.1</td>
<td>0.0</td>
<td>0.1</td>
<td>0.2</td>
</tr>
<tr>
<td>MgO</td>
<td>18.3</td>
<td>20.8</td>
<td>8.1</td>
<td>4.6</td>
</tr>
<tr>
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<td>6.8</td>
<td>38.6</td>
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<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>K₂O</td>
<td>6.2</td>
<td>7.6</td>
<td>1.6</td>
<td>0.2</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>4.2</td>
<td>4.1</td>
<td>4.5</td>
<td>0.5</td>
</tr>
</tbody>
</table>

¹) Average ore composition; significant variation from block to block. ²) Mainly tetraferriphlogopite.
Several younger intrusions crosscut the entire Siilinjärvi complex. Northwest–southeast or NNW-SSE trending mafic dykes, with widths that range from few centimetres to several tens of metres, transect the ore. The southwestern margin of the main ore body has been intruded by mafic dykes and bodies of Paleoproterozoic tonalite-diorite displaying mingled texture. These are associated with disruption and N-S shearing of carbonatite-glimmerite ore (O’Brien et al. 2015).

The Siilinjärvi mine is currently the only operating phosphorus mine in Western Europe, the closest competing operator in the Arctic being in the Kola alkaline province of the northwest Russia. The Siilinjärvi complex was discovered in the 1950 after a local mineral collector sent a sample of carbonatite to the Geological Survey of Finland (GTK).

The Sokli intrusion (Figure 33) is a multi-stage, funnel-shaped pluton ca. 5 km in diameter that intrudes the local Archean crust. The intrusion consists of a magmatic carbonatite core concentrically surrounded by metacarbonatite, metamorphically altered ultramafic rocks, and a wide fenite aureole (Vartiainen 1980, 2001; O’Brien et al. 2005). Late-stage carbonatite dikes cross-cut the older intrusion phases and the fenite zone. The intrusion is capped by a ca. 26 m thick regolith that constitutes the main phosphate ore. In addition to phosphate ore the deposit contains Nb, Ta, Zr and some U. The main ore minerals at Sokli are apatite, pyrochlore, magnetite, ancyrite, rhabdophane, baddeleyite, zirconolite, fersmite and lueshite. The Sokli carbonatite complex is estimated to contain in total 190.6 Mt of P ore grading 11.2 wt.% P$_2$O$_5$. The ore comprises lateritic P ore (36.7 Mt @ 18.7 wt.% P$_2$O$_5$), silicate-apatite P ore (19.3 Mt @ 11.0 wt.% P$_2$O$_5$), weathered crust P ore (57.6 Mt @ 14.1 wt.% P$_2$O$_5$), and Kaulusrova ore (75.2 Mt @ 5.6 wt.% P$_2$O$_5$, 0.2 wt.% Nb$_2$O$_5$ and 1.9 Mt Nb-Ta ore @ 1.9 wt.% P$_2$O$_5$, 0.6 wt.% Nb$_2$O$_5$), which are all considered as soft rock ores (Siirama 2009, Pöyry Environment Oy 2009). Siirama (2009) also reports hard rock ores that comprise 3000 Mt in the magmatic core, 110 Mt Nb ores, 500 Mt Arch ore (situated on the SE side of the carbonatite core), 1800 Mt of metacarbonatic ores and 6800 Mt of metasomatite ores, which altogether yield a total of 12 210 Mt of P-ores in hard rock. The so far unexploited phosphate deposit is currently under a mining license held by Yara Suomi Oy. Recent investigations by GTK indicate that late-stage carbonatite ring dykes surrounding the intrusive complex contain REE carbonates such as ancyrite-(Ce) and bastnäsite-(Ce). The Sokli project of Yara Suomi was suspended in autumn 2015 due to economic reasons.
Figure 33. Geological map of the Sokli carbonatite complex (modified after Vartiainen 1980). Legend: 1-central fracture zone, 2-magmatic phosphorite, 3-metaphosphorite, 4-magmatic sövite and silicosövite, 5-metasilicosövite, 6-metasomatite, 7-fenite, 8-amphibolite, 9-tonalite gneiss.
Kimberlites and related rocks occur in three major provinces in eastern Finland, (1) Kaavi-Kuopio, (2) Lentiira-Kuhmo and (3) Hossa-Kuusamo. Provinces 1 and 3 contain archetypical kimberlites, whereas only olivine lamproite-orangeite hybrids are known from province 2. Altogether, there are approximately 40 known kimberlitic/lamproitic bodies in Finland. Over half of the bodies are proven to contain diamonds; mostly, however, only microdiamonds. The most diamondiferous kimberlite found so far, is the Lahtojoki Kimberlite pipe that belongs to the Kaavi-Kuopio province.

The Kaavi–Kuopio province consists of two kimberlite fields, located only ~ 50 and ~ 30 km from the southwestern margin of the Karelian craton. The kimberlites are characterised by abundant macrocrysts of olivine, in a matrix of euhedral olivine, monticellite, perovskite, magnesian ulvöspinel-magnetite, Ba-rich phlogopite-kinoshitalite mica, apatite, calcite, and serpentine (O’Brien & Tyni 1999). Their major and trace element contents as well as Sr, Nd, and Pb isotopic compositions are similar to those of other Group I kimberlites globally. Age determinations by ion microprobe of U-Pb in perovskites from four pipes have resulted in ages ranging from 589–626 Ma (O’Brien et al. 2005).

The Kaavi-Kuopio kimberlites range from hypabyssal dikes to volcaniclastic kimberlite and volcaniclastic kimberlite breccias formed in diatremes. The bodies are rather small, ranging from < 1 m dykes to diatreme-facies pipes up to 4 ha in surface area. None of the pipes have crater-facies materials preserved due to erosion. Many of the pipes are diamondiferous, several having reasonable diamond grades (14 – 41 ct/100 t; Tyni 1997).

The Lahtojoki Kimberlite is the best studied of all the kimberlites due to its high diamond content. It is located in Kaavi municipality (63° 1’ 50.3"N, 28° 34’ 39.2"E). The pipe was discovered by Malmikaivos Oy by sampling of glaciogenic deposits followed by magnetic survey and drilling in 1989. The Lahtojoki diatreme is suboval in plan, measuring approximately 200 m (E–W) × 100 m (N–S), with approximately 2 ha of surface area. It is located in a swamp and covered by 13–20 m of glacial deposits and peat. Drill core information suggests that at the −100 contour, the pipe plunges 80° toward the south, and is still nearly 2 ha in cross section.

The Lahtojoki kimberlite is mainly composed of macrocrystal tuffisitic kimberlite and subordinate tuffisitic kimberlite breccia with relatively rare hybabysal kimberlite. The degree of weathering varies in the upper part of the pipe. Locally there is a several metres thick soft weathered kimberlite horizon whereas in other areas the weathered kimberlite has been removed by glacial erosion exposing fresh kimberlite.

Mini-bulk sampling of the pipe by Malmikaivos Oy/Ashton Mining Ltd, totalling at 23.3 t, suggests an average grade of 26 carats/100 t for diamonds > 0.8 mm in diameter (Tyni 1997). However, the 1000 t bulk sample collected by the company returned only 5.7 c/100 t. The large discrepancy between the results is probably due to the high country rock dilution of the volcaniclastic kimberlite breccia that formed a significant portion of the bulk sample. In contrast, the volcaniclastic diatreme material that was sampled for mini bulk testing, has a much lower country rock content overall. The notable variations within the mini-bulk results may, however, be related to the fact that diamonds in the Lahtojoki kimberlite occur not only as xenocrysts in the matrix, but also as a rock-forming mineral in some of the eclogite xenoliths (Peltonen et al. 2002).
Figure 34. Volcaniclastic kimberlite breccia from the Lahtojoki kimberlite. Photo: Niko Auvinen, GTK.

Figure 35. An eclogite xenolith from Lahtojoki hosting a diamond crystal. The diameter of the diamond is 1 mm. Photo: Kari A. Kinnunen, GTK.
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Deposits
- Diamonds
- Hydrothermal fields
- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferrous metals: Cr, Fe, Mn, Ti, V

Size and activity
- Active mine
- Important deposit
- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Scale 1:23 000 000
Stereographic North Pole Projection
Standard Parallel 70°N Coordinate System WGS 1984
Prime Meridian: Greenwich (0.0), Central Meridian 20°W
CHAPTER 9
RUSSIA


A.P. Karpinsky Russian Geological Research Institute (VSEGEI)
OUTLINE OF THE GEOLOGY OF RUSSIA
NORTH OF 60°N

This outline has been compiled, on the basis of a number of publications from the last six years, in order to provide a frame of reference for the following sub-chapters which focus on groups of commodities.

The Murmansk and Karelia regions

The north-westernmost part of Russia, consists mainly of Archaean and Proterozoic rocks, divided into three main provinces – the Kola, Belomorian and Karelian provinces (Figure 1): the Rybachy Peninsula, on the north coast of the Kola Peninsula, consists of a sedimentary sequence deformed during the Neoproterozoic Timanide Orogen which has its continuation west of the northern Urals.

The Kola Province

The Kola Province has two sub-provinces, the coastal Murmansk sub-province and the Kola Province, in the specific sense, which occupies most of the Kola Peninsula.

The Murmansk sub-province (Hölttä et al., 2008) consists of Neoarchaean granitoids, diorites, enderbites and a lesser component of supracrustal rocks: the effect of Palaeoproterozoic deformation and metamorphism is not extensive.

The Kola Province (Hölttä et al., 2008) encompasses four terranes:

- The Kola-Norwegian terrane in which Neoarchaean tonalite-trondhjemite-granodiorite (TTG), diorites, enderbites and peraluminous metasediments are prominent. The terrane contains several volcanic belts including the Pechenga volcanic/intrusive suite, which is host to important nickel-copper ores.
- The Kolmzero-Voron’ya terrane, which is a Mesoarchaean suture zone with components of komatiite, arc-related tholeiitic basalts, andesites and dacites as well as conglomerates. The terrane also includes Neoarchaean monzodioritic and granitic intrusives.
- The Keivy Terrane encompasses Mesoarchaean continental-margin volcanic suites, a suite of alkali granitoids, gabbro-anorthosite intrusives and extensive units of coarse-grained kyanite schist.
- The Sosnovka terrane, close to the south-eastern coast of the peninsula, consists of TTG rocks.

The Devonian alkaline intrusive complexes, Khibiny and Lovozero, are prominent features in the central part of the Province, forming two of the largest alkaline complexes in the world and containing rich and very large resources of special metals.

The Belomorian Province

This province consists (Hölttä et al., 2008) mainly of TTG gneisses, greenstones and paragneisses of Mesao- and Neoarchaean age, deformed and metamorphosed in Neoarchaean and Palaeoproterozoic orogenic events. It is also noteworthy in that it contains ophiolites and eclogites.

The Karelian Province

This province consists (Slabunov et al., 2006) of three terranes:

- The Western Karelian Terrane comprises part of Eastern Finland and the western part of Russian Karelia. The most common lithologies (Hölttä et al., 2008) are migmatisic TTG orthogneisses and amphibolites, ranging in age from Palaeo- to Neoarchaean. It also includes Mesoarchaeane greenstone belts, including the Kostomuksha belt which hosts important banded iron formations.
- The Central Karelian Terrane contains
intrusions varying from granitoids to felsic to ultramafic sanukitoids, as well as several greenstone belts. These are, unlike the lithologies in the neighbouring terranes, exclusively Neoarchaean.

- The Vodlozero Terrane in the southeastern part of the Province has a core of Palaeoarchaean granitoids and gneisses, intruded by Mesoarchaean granites and mafic complexes, and surrounded by three generations of greenstone belts, ranging in age from Meso- to Neoarchaean.

The East European Platform

The whole northern coastline of European Russia, from the Kanin Peninsula, east of the opening of the White Sea to the Ural Foredeep (see next section) consists of the rocks of the Timanide Orogen (see Figure 2, unit 10), which form a wedge-shaped exposure, the southwest margin of which meets the western margin of the Ural Foredeep at ca. 60°N. South of the Timanides the platform cover is represented by two different complexes (www.rusnature.info):

- Riphean to lower Vendian sediments deposited mostly within deep basins (grabens, rifts or aulacogens) and consisting of terrigenous sandy-clayey rocks with a thickness of up to 5 km, locally with units of basaltic rocks.
- Late Vendian to Phanerozoic sediments forming vast synclines and basins with a thickness varying from tens of metres to 20-22 km in the deepest basins.

The Ural Mountain Chain

The Ural mountain chain extends for ca. 2500 km from the Aral Sea in the south to the Barents Sea in the north, with a continuation on Novaya Zemlya north of the Kara Strait. The province is
a fold-and-thrust belt including the Pre-Uralide sub-province to the west and the main Ural sub-province: the Urals are bounded to the west by the East-European Platform (overlain by the Timan-Pechora Basin) and to the east by the West Siberian Platform and overlying Mesozoic-Cainozoic sediments.

The geological development of the province consists of the following stages:

- Pre-Uralian (Archaean – Early Proterozoic)
- Baikalian or Early Uralian (Riphean – Vendian)
- Uralian – Late Uralian (Ordovician-Permian)
- Development of platform cover (Mesozoic-Cainozoic)
The geotectonic development of the belt includes: subduction-island-arc, collisional, platform and rift-related settings, each with particular, though not equally important metallogenic characteristics. The longitudinal zoning of the province, from west to east is, as shown in Figure 3:

- **Uralian Foredeep**: Permian molasse
- **West Uralian Zone**: W-directed nappes of Palaeozoic sedimentary sequences
- **Central Uralian Zone**: exhumed Precambrian complexes
- **Main Uralian Fault Zone**
- **Magnitogorsk-Tagil Synclinorium (MTS)**: Palaeozoic ophiolite and island-arc complexes
- **East Uralian Upland**: as MTS but also includes Precambrian complexes
- **Trans Uralian Zone**: pre-Carboniferous, probably accretionary complexes, overlain by L. Carboniferous volcanic rocks.

The longitudinal tectonic zones are cut by NW trending transverse structures which split the belt into four megablocks (Southern, Middle, North and Polar Urals) characterized by specific geological, tectonic and metallogenic features. The presence of multiple structural levels and the complex tectonic zoning of the belt necessitate the definition of a system of longitudinal and transversal metallogenic zones with different ore-geochemical features. The main resource-types found in the metallogenically most important zones (Puchkov, 2016) are:

- **West Uralian Zone**: barytes, sediment-hosted Cu-Zn.
- **Central Uralian Upland**: titanomagnetite, iron, chromite, gold, VMS deposits, Mo-W in granite and others.
- **Magnitogorsk-Tagil Synclinorium**: VMS deposits, Cu and Au-Cu porphyry, platinum metals.

**Siberia**

Siberia incorporates several large tectonic terranes (Seltmann et al., 2010).

- The Siberian craton
- The Western Siberian Lowland
- Neoproterozoic–Palaeozoic orogenic belts north and south of the craton
- Late Palaeozoic–Mesozoic orogenic belts east of the craton

The following description is based on that of Seltmann et al. (2010), focusing on the area north of 60°N.

**Siberian craton**

The Siberian craton extends from the R. Yenisei in the west to the R. Lena in the east. The Archaean basement of the craton is exposed in the Anabar and Aldan Shields and in smaller uplifted blocks along the margins of the craton. The platform cover on the craton includes sedimentary and volcanic sequences of Mesoproterozoic, Vendian–Cambrian, Palaeozoic, Mesozoic and Cainozoic age. The cover sequences are intruded by several igneous suites, including the voluminous Siberian traps and related mafic/ultramafic intrusives, ultramafic–alkaline rocks, carbonatites and kimberlites.
Anabar and Aldan Shields

The Anabar and Aldan Shields are the largest uplifts of ancient basement within the Siberian craton (Figure 4). The Anabar Shield is composed of Archaean and Proterozoic granulites overthrust by Early Proterozoic granite–greenstone terranes. The Aldan Shield is also composed of Archaean and Proterozoic metamorphic sequences, including major granitoid intrusives, greenstone belts and sedimentary sequences. The Archaean and Proterozoic rocks are discordantly overlain by subhorizontal Mesoproterozoic–Vendian (750–550 Ma) and Cambrian carbonate sequences that represent cratonic cover.

Proterozoic terranes of the Yenisei Ridge and Patom Highlands

These two orogenic belts, adjoin the western (south-western) and southern (south-eastern) edges of Siberian craton, respectively (Baikalides in Figure 4). They include Archaean terranes (crystalline schists, granulites, gneisses) and predomiating thick Mesoproterozoic–Neoproterozoic sequences. The latter are composed of metamorphosed shales, sandstones, siltstone, and minor marbles, dolomites and amphibolites that formed on passive continental margins (Obolenskyi et al. 1999). Precambrian granitic and mafic alkaline rocks are locally found. The Precambrian sequences are overlain, locally, by a Lower Cambrian carbonate sequence.

Western Siberian Lowland

The Western Siberian Lowland (Figure 4) is comparable in size to the Siberian craton and underlies most of Siberia west of the craton and is bordered by the R. Yenisei to the east and the Urals to the west. A large orogenic system referred to as the Yenisei Ridge separates southern parts of the Western Siberian Lowland from the Siberian craton. The basement of the Western Siberian Lowland immediately east of the Urals consists of Precambrian and Palaeozoic folded structures of the Altai–Sayan orogenic belts. The basement is overlain, in most of Western Siberia, by 4–6 km of terrigenous Mesozoic–Cainozoic sediments.

Neoproterozoic–Palaeozoic orogenic belts of Northern Siberia

The Taimyr–Novaya Zemlya orogenic belt adjoins the Siberian craton in northwestern Siberia. This terrane is separated from the craton by the Yenisei–Khatanga trough which is composed of Triassic–Oligocene marine–terrigenous sedimentary rocks, and has a Baikalian (Neoproterozoic–Mesoproterozoic) metamorphic basement, which contains blocks of Archaean age (micro-continents). This basement is bordered to the north by Mesoproterozoic–Devonian terrigenous–carbonate rocks and to the south by Ordovician–Triassic terrigenous-carbonate rocks with minor basalts. Orogenic belts in Southern Siberia are part of the Central Asian Orogenic Belt or the Altaid orogenic collage.
Yana–Chukotka orogenic belt
This Mesozoic orogenic belt extends from the northeastern margin of the craton to the Pacific Ocean (Mesozoides in Figure 4). It consists of several terranes/orogenic belts, including:

- The Verkhoyansk foreland and anticlinorium immediately east of the craton,
- The Yana–Kolyma thrust and fold belt,
- The Kolyma and Omolon microcontinents and
- The Oloy and Chukotka terranes.

Formation of this superbelt began on the eastern passive continental margin of the Siberian craton in the Mesoproterozoic and culminated in the Permian–Early Jurassic with deposition of thick sandstone–shale sequences. In the Late Jurassic–Cretaceous, this succession was folded, faulted and intruded by granitic batholiths. These rocks now form the Verkhoyansk and Yana–Kolyma terranes. The Kolyma microcontinent separates the Yana–Kolyma thrust and fold belt from the Oloy and Chukotka terranes: it consists of Archaean and Proterozoic rocks surrounded by Ordovician–Carboniferous sequences, mainly consisting of carbonate rocks. The superbelt incorporates, in addition to the Kolyma microcontinent, several smaller continental blocks, including the Omolon, Okhotsk, Taiganoss and Chukotka microcontinents: these consist of Palaeo-proterozoic schists overlain by Mesoproterozoic, Palaeozoic and Mesozoic sedimentary successions.

Okhotsk–Chukotka volcanic belt
This Mesozoic–Cainozoic volcanic belt overlies the eastern part of the Yana–Chukotka orogenic superbelt and thus separates the latter from the Cainozoic Koryak– Kamchatka orogen to the east (Figure 4). The Okhotsk–Chukotka volcanic belt thus has a Mesozoic basement of Upper Triassic to Lower Cretaceous sedimentary and volcanic sequences and is unconformably overlain by Lower Cretaceous to Palaeocene volcanic sequences, locally 45 km thick. The latter include rhyolite (ignimbrite), andesite and basalt. The Okhotsk–Chukotka volcanic belt evolves, further south along the Pacific coast, into the Sikhote–Alin orogen (Figure 4), which is characterised by volcanic rocks including Late Cretaceous rhyolites, Late Cretaceous–Palaeogene rhyolite–dacite, Late Palaeogene rhyolite–basalt–andesite suites and Neogene andesite and basalt.

Intraplate (or intracontinental) tectonic and magmatic activity
The Siberian terranes have experienced several intense intraplate anorogenic tectono-thermal events. These include the intrusion of igneous suites such as the Siberian traps, alkaline and ultramafic-alkaline complexes, carbonatites, kimberlites and lamproites. The Siberian traps have an extent of ca. 7100 km², including continental low-Ti tholeiitic flood-basalts, their possible feeders and comagmatic sill-like intrusions (Ivanov et al. 2008). The giant Noril’sk and Talnakh nickel–copper–PGM deposits are hosted in mafic sills associated with the traps. Intraplate tectonic and magmatic processes also resulted in emplacement of kimberlites in Yakutia, many of them diamondiferous, between the Devonian and the Mesozoic.

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BRIEF HISTORY OF MINING IN RUSSIA

18th – 19th Centuries

Tsar Peter the Great (1672–1725), was responsible for numerous initiatives in Russia, also in relation to the search for and exploitation of mineral resources (http://goldminershq.com/vlad.htm).

One of the first important mines was at the Nерchinsk silver deposit, southeast of Lake Baikal, which was discovered in 1702: the mines in the district were in operation from 1704 to 1854, yielding 11,540,000 oz (327,154 kg) of silver. Placer gold was discovered in the area in 1830 and was an important product for the last period of operation.

Primary gold mineralizations were discovered on the north coast of the White Sea in 1937 and in the eastern Urals in 1745, the latter leading to a mining operation from 1748: the number of discoveries in the latter area by 1800 reached ca. 140, several of them in production as part of a state monopoly. The state monopoly was disbanded in 1812, leading to a dramatic increase in exploration activity and alluvial gold mining, especially in the Urals, and, in the second half of the 19th Century in eastern Siberia: Russia, has, thereafter, been one of the world’s major producers of gold.

Numerous discoveries were made and mines established in Karelia in the first half of the 18th C, among the most important being Voitskoe (Cu) and the Pitkäranta (Cu-Sn-Fe) (Eilu et al., 2012). The mining industry had, by the late 19thC, evolved dramatically, to include major production of Au, Ag, Pt, Cu and Fe in the Urals, Fe and Mn in southern Russia (now the Ukraine), Pb-Zn, Ag and Cu in the Caucasus and Au and Ag in Siberia (https://en.wikipedia.org/wiki/Russian_Empire).

20th Century

The mining industry continued its expansion until the period prior to World War I and the Revolution. The "World War, the civil struggles and foreign invasions" caused a dramatic decline in industrial production in general in the period 1913 – 1921-22 (https://www.marxists.org/history/ussr/government/1928/sufds/ch05.htm). Production of certain metals fell to below 10 % of pre-World War I levels and did not recover until the period of dramatic expansion of the mining industry in the 1930s which continued for decades after World War II. The following sub-chapters indicate implicitly, in the descriptions of numerous major deposits, the efforts made in exploration for and documentation of new deposits of many commodities, several of which are among the largest of their kind in the world. Exact data on the reserves documented, and on production levels, both at the deposit and commodity levels are not publicly available for past discoveries. Current practice, however, allows publication of data on reserves and resources and their grades for deposits of most commodities. Russia was, in 2014, the world’s most important producer of diamonds and palladium and one of the three largest producers of antimony, gold, nickel, platinum, tungsten and vanadium (BGS, 2016).

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GOLD, SILVER

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Blagodatnoe gold deposit

The Blagodatnoe deposit (not to be confused with the similarly named deposit in Khabarovsk krai) is located in the Proterozoic foldbelt on the northwestern margin of the Siberian Platform. It is located on the SW flank of the Panimbinsky Anticlinorium, which hosts most of the gold deposits of the Yenisei Ridge. The deposit forms a linear (NW strike and NE dip) vein-and-veinlet, sulphide-gold-bearing zone of hydrothermally altered (silicified, sericitized, and carbonatized) Upper Proterozoic quartz-mica schist of the Gorbiloksky Formation (Sovmen et al., 2006) (according to other data, schist of the Kordinsky Formation). Two commercial gold ore bodies have been identified in a 100-400 m thick zone, one in the northern prospect and the other in the southern prospect (Figure 5). The total length of the ore bodies is 3250 m; their thickness varies from 5-148 m, 45.6 m on average. The ores are low-sulphide, veinlet-impregnated and contain no commercial components other than gold. The structure of the ore is spotted, banded – the veinlets are banded and plicated. The host rock is dominated by quartz, feldspar, mica and chlorite: garnet, carbonate, whereas sulphides are subordinate. The main sulphides are arsenopyrite, loellingite, pyrrhotite, marcasite and pyrite: chalcopyrite, galena and sphalerite are less common. Rammelsbergite, nickeline, tellurides (hessite, tellurobismuthite and altaite), sulphosalts, sulphoarsenide and sulphoantimonide are also recorded. The purity of the gold varies from 67–97 %; 78–90 % on average. The dimensions of the gold grains are typically between 10 and 750 µm, but grains of up to 2-3 mm have been found. Most of the commercial gold is confined to gold-sulphide intergrowths among non-metallic minerals. Most of the fine-grained gold (up to 70 µm) is concentrated in the arsenic-bearing phases.

The Blagodatnoe deposit was discovered in 1968. The inferred resources of the deposit were assessed, in 1975, to contain 40 tons of gold. In 2000-2004, an exploration crew from ZAO Polyus made an assessment of the ore within the framework of a project for exploration and appraisal in the Olimpiadinsky area. As a result, the Blagodatnoe ore showing (northern prospect) was reevaluated and a new, southern prospect that hosts 4/5 of all reserves of the deposit, was discovered. In September 2005 reasonably assured and inferred reserves containing 222.4 tonnes Au (ave. grade: 2.4 ppm Au) were approved within the pit planned in the Blagodatnoe deposit. The gold processing plant was put into operation in 2010. The mine is operated by ZAO Polyus (a subsidiary of OAO Polyus-Zoloto). The gold is extracted by direct leaching, and in 2010 76.73 % of the gold in the ore was recovered. The gold production in 2009 was 1.3 t Au, but increased to 11 to 13 t in the following years (2010-2013). Reserves as of 01.01.2012 were: RAR – 250.5 t Au, with an average gold grade in the ore of 2.4 ppm; inferred resources – 34.7 t Au.

Mayskoye gold mine

The Mayskoye mine is located in the Chaun District of the Chukotka Autonomous Region, 150 km SE of the regional centre, Pevek. The ore field was discovered by S.A. Grigorov during the 1971-1972 geological survey in the Tamnekvun tin ore cluster (Kukenev horst and
its surroundings). Anomalous gold mineralization was identified over an area of about 2 km² in the "Pakovlad" prospect (now Mayskoye mine). The discoveries were followed up by a detailed survey and assessment in 1973 by the Maysky Production Geological Enterprise of the Chaun Complex Exploration Expedition.

The Mayskoye mine is in the upper reaches of the Keveem River in the eastern Ichuveem-Palyavaam ore district, in which synvolcanic magma-feeding deep faults of NE and near N-S strike occupy an oblique position with respect to the NW striking cofolded faults. The main control on ore location is its confinement to the domeblock structure, located close to the Kukeney intrusion of Early to Late Cretaceous age (Figure 6).

The ore field area of 10 km² has an isometric shape and is confined to a complex horst-like ledge located at the intersection of the NW-, NE- and approximately E-W- and N-S-trending faults. The host rocks are interbedded mudstones, siltstones, and oligomictic sandstones of the Middle Triassic, Karnian, and Lower Norian (Keveem, Vatapaam, Relkuveem, and Melyuveem formations). The host rocks are sandy-silty-shale deposits of the Keveem Formation of the Middle Triassic. Numerous pyrite nodules are observed in shaly silt varieties of the formation. The formation is, in the central part of the mine, at least 600 m thick. The folded structure of the ore field comprises gently dipping fault zones and folds alternating with zones of vertical, steeply-dipping isoclinal folds.

Thus, multidirectional folding, combined with numerous intersecting faults, has created the mosaic structure of the ore cluster. Igneous rocks occupy 25% of the total volume and are represented by the Early-Late Cretaceous dykes. Their outcrops form a belt about 3 km wide and more than 4 km long, spatially associated with mineralized zones. Two dyke groups of different age have been distinguished. (Volkov, 2005). The first group includes granite-granodiorite-porphyry, aplite, and lamprophyre, the second, later sub-volcanic rhyolite porphyry. It is important to note that both the igneous and the sedimentary rocks, have undergone intense epigenetic processes. Quartz is recrystallized and has reaction rims of sericitic composition; feldspar is commonly totally replaced by sericite, chlorite, and clay minerals of the kaolinite-dickite group.

The Mayskoye mine is divided into three tectonic units: the Western, Central, and Eastern units (Figure 7), in which dykes and ore bodies are combined into eight near N-S-oriented ore zones 100 - 300 m wide and 300 - 2,500 m long. Forty-six ore bodies are distinguished in the mine, 28 of them with officially reported ("balance") reserves. The ore bodies are represented by near N-S trending zones of cataclasites 1 - 8 m thick, composed of altered rocks. Up to 85% of the gold reserves are concentrated in ore body No.1 of the Central unit; 90% of the ore is refractory, with finely dispersed gold and requires a complex...
enrichment technology. The Mayskoye mine is among the large ones, with gold reserves of over 200 tons. Ore bodies of the Mayskoye mine are represented by sulphidized linear zones of cataclasites and have an approximately N-S trend. They form a system of near E-W bodies in an E-W-striking zone, 3.5 km long and limited to the north and south by E-W striking faults. Over 30 ore bodies are registered, most of them not exposed at the surface. The ore body boundaries are indistinct due to the development of mineralization on shallow feathering fractures and cleavage zones in host rocks.

The ore bodies are composed of intensely broken and crushed mineralized rocks: silt-stone, shale, fine-grained sandstone, rarely rhyolite porphyry containing 6-10 % of fine-grained, crystalline disseminated sulphides (pyrite, pyrrhotite, arsenopyrite and stibnite). Cataclastic structures, linear breccias including sedimentary and igneous rocks cemented by sulphides and quartz are common. In general, the average thickness of the ore bodies at the mine is 2 m, the average content of gold is 12 g/t, antimony 0.25 %, carbonaceous matter 0.5 %, arsenic 1 % and silver 3 g/t. Over 90 % of the gold is finely dispersed and associated with sulphides, mostly with arsenopyrite, arsenic-bearing pyrite and pyrrhotite (Figure 8). Significant positive correlation is observed between gold and arsenic. There are, in addition, areas with native gold in the fragmented parts of primary quartz veins. The gold is high-grade (920-980 ‰), forming a solid mass cementing quartz fragments (Figure 8). Currently, the license for production of gold and associated components at the Mayskoye gold mine and for geological study of its flanks and deep horizons belongs to LLC "ZK" Mayskoye", a subsidiary of the "Polymetal" holding company. Mining started in December 2012, and in March 2013, the gold processing plant produced its first concentrate. The first metric ton of gold from the mine was produced in October, 2013.

**Kubaka gold mine**

The Kubaka mine is located in the North Evenk District of the Magadan Region in the Kolyma massif, 1,000 km from Magadan. The ore field

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**Figure 7:** Geology of the Mayskoye ore field (Grigorov 1980).
1) Vatapvaam Formation (Lower); 2) Vatapvaam Formation (Upper); 3) Relkuvem and Mlelyuveem formations; 4) and 5) Quartz-porphyry and rhyolite dykes; 6) Lamprophyre dykes; 7) Quartz-feldspar porphyry dykes; 8) Fault zones (at surface and blind); 9) Ore bodies (at surface and blind); 10) Lithological contacts.
Figure 8: Quartz-antimonite vein with kaolinite and native gold. From the Mayskoye deposit.
was discovered in 1979 by geologists of the Seymchan Prospecting Expedition. Exploration to determine the technological properties of the ores, their position and the morphology of the ore bodies was carried out from 1979 to 1992.

The Avladinsky ore cluster, including the Kubaka mine, is located in the Kedon rise of the Omolon massif. The Precambrian of the Kedon rise is characterized by the presence of both normal leuco- and mesocratic gneiss of

amphibolite facies and pyroxene-bearing mafic rocks of granulite facies (including clino- and orthopyroxene). The ore cluster is located in Middle-Late Devonian Kedon Group volcanics and coincides spatially with a palaeovolcanic structure with the same name. The volcanic structure has vent, nearvent, and peripheral parts, separated by concentric fractures. The vent part is composed of ignimbrites and clastic lavas of trachyrhyolitic composition: tephroids with foamy rhyolite lava flows are observed locally. The periphery of the volcanic structure consists mainly of distal facies - tuff-sandstone, tuff-siltstone, acidic and intermediate tuffs including ash and silica tuff, rare ignimbrites and agglomerate tuffs. A number of sub-volcanic bodies of trachyrhyolite and rhyodacite occur, mainly intruded into concentric fractures.

Figure 10: Geology of part of the Central ore zone from the Kubaka mine and section (Konstantinov et al, 2000). 1: Agglomerate tuff; 2: Rhyodacitic ignimbrites; 3: Rhyodacite; 4: Tuff-sandstone; 5: Adularia-quartz veins: a – of the first producing stage, b – of the second producing stage; 6: Wallrock alteration of chlorite-quartz-sericite and quartz-sericite facies; 7: Pre-ore propylite of epidote-chlorite and carbonate-chlorite facies.
Alteration of volcanic rocks includes propylitization, and large fields of altered rocks are developed on the flanks of the volcanic structure. Altered rocks are accompanied by gold-bearing vein-veinlet zones consisting of adularia-quartz, quartz, and carbonate-quartz. Gold mineralization, accompanied by "aureoles" of alluvial gold, occurs unevenly distributed in the outer, peripheral part of the volcanic structure. The mineralization is concentrated in two relatively small mineralized areas, 30-35 km² in size, limited by concentric and radial faults, on the southwestern and northeastern flanks of the volcanic structure. These areas are called, respectively, the Kubaka and Strelinsky-Gruntovsky ore fields.

The known ore bodies in the mine are concentrated in an area of about 8 km² which is elongated in a NW direction (Figure 9). The ore-bearing unit is composed of stratified volcanics of the Kedon Group, forming a monocline dipping gently to the southwest (200-210°) at an angle of 10-15°. Rhythmically alternating effusive rocks of intermediate and felsic compositions are observed. The volcanic rocks are intruded by sills of rhyodacite and andesite-dacite, as well as by trachyrhyolite dykes of the same age and later dolerite dykes. The southern flank of the ore body is covered by unconformably bedded terrigenous sediments of the Early Carboniferous Korbin Formation.

The mineralization is associated with veinlet-vein zones localized within a system of intersecting steep faults oriented to the northwest. Eighty-five percent of the main gold reserves are concentrated in the Central zone, 10 % in the Socle zone, and 5 % in the Northern zone. The veinlet-vein zones are separated into a number of ore bodies by a system of strike-slip faults. The faults are characterized by multiple brecciation and superposition of several stages of mineralization.

A system of E-W-trending composite veins of late-stage mineralization is always present in selvages of the veinlet-vein zones. In cross-section the ore zones form a fan-shaped structure widening upwards. The veinlet-vein zones are accompanied by hydrothermally altered rocks of chlorite-hydromica facies, forming an aureole up to several tens of metres wide. The thickness of the mineralized veinlet-vein zones ranges from a few metres to 20-30 m: the composite veins are 1-3 m wide.

The gold-silver mineralization is mainly localized in the marginal parts of the veins (Figure 10). Vein systems can be traced to depths of 500-700 m. Commercial mineralization is localized in the upper parts of the Kedon Group volcanic rocks; its vertical depth reaches 220 m. The ore is localized in an early stage of carbonate-anthothoclase-quartz mineralization and a late stage of chlorite-adularia-quartz mineralization, separated by the formation of weakly mineralized breccias with chalcedony-quartz cement. Hydro-muscovite is also observed in the veins, and ore channel No. 1 contains fluorite and barytes. The ore is dominated by pyrite (50-80 %). Other phases are gold, electrum, custerite, acanthite, hessite, arsenopyrite, galena, sphalerite, hematite, ilmenite, ilmenorutile, rutile, chalcopyrite, freibergite, naumannite, aguilarte, native silver and other, rare minerals. The gold-silver ratio in the ores is high (1:2 - 1:1). Gold fineness is 600-750 ‰. Impurities include mercury (up to 1.48 %), antimony (up to 1.4 %), selenium and tellurium. The gold is mainly hosted in arsenopyrite and pyrite. Hypergene fine gold occurs very rarely in intergrowths with iron hydroxides. Balanced (officially reported) reserves as of 01.01.2013 were: gold: A+B+C1 – 9.031 t, C2 – 5.009 t.

**Dvoynoye gold mine**

The Dvoynoye mine is located 140 km SE of the town of Bilibino and 330 km SW of the town of Pevek and connected to them by a seasonal winter road. The nearest community is Ilirney village. The commercial ore bodies of the Dvoynoye mine (zone 1) were found by the Anyui State Mining and Geological Enterprise in 1984-1985 during a geological survey at 1:50,000 scale. Further exploration was carried out in 1986-1988. The mineralization of zone 1 was traced along strike and dip, as well as ore zones 8, 37 and 38.

The Dvoynoye mine is located on the southeastern flank of the Anyui folded zone near its junction with the Okhotsk-Chukotka volcanicogenic belt. The ore field has an area of about 20 km² and consists of andesite and andesitic tuff of Aptian-Albian age (Figure 11). Intrusive formations are represented by subvolcanic liparite bodies, granitoids of the Ilirney heterogeneous Early Cretaceous massif and Late Cretaceous dykes of diorite porphyry and andesite-basalt.
Thirteen mineralized vein-veinlet zones have been identified within the ore field, 4 of them being of economic interest and which have been explored in varying detail. Ore zone 1, in which all the officially reported reserves of gold and silver are found, has been exploited by underground mining.

Ore zone 1 is confined to the exocontacts of a granite-porphyry dyke and has a total length of 1,400 m. The length of the productive part is 400 m and its thickness is 25-30 m. The vertical extent of the mineralization is 310 m and has been explored to a depth of 100 m. The zone has a NW strike with a steep (70-80°) dip to the northeast (Figure 12). The major gold and silver mineralization is concentrated in the upper exo-contacts of the porphyry dyke. Morphologically, the exo-contact zone is divided into three segments. The northwestern and southeastern segments are ore shoots. The individual veins within these segments are up to 8 m thick and the total thickness of the segments reaches 23 m. The central part of zone 1 is confined to pinching part of the porphyry dyke and is represented by a zone of veinlets up to 2.5 m in thickness.

Ore zone 37 is up to 100 m thick and consists of closely spaced individual quartz vein bodies, veinlet zones and zones of altered host rocks. The ore zone is exposed on the banks of the Dvoynaya River. The zone is overlain by a post-ore, sill-like, complex intrusive subvolcanic body of syenite, quartz syenite and rhyolite along most of its southern part. In the northern continuation the zone is covered by felsic volcanics. Along strike the zone can be traced for 940 m, down-dip to a depth of 320 m from the surface. The thickness of quartz and quartz-adularia veins varies from 0.1 m to 42 m, over lengths from 45 m to 590 m. Many vein systems are not persistent along strike and down-dip: they have lenticular, en-echelon structures and branch both down-dip and along the strike. Among the numerous quartz vein systems, 13 quartz veins with elevated gold and silver contents have been identified to date. There is a great variety of ore textures: stringer, breccia, framework-tabular, crustified, massive, banded, etc. The quartz vein systems consist of quartz (80-90%), adularia (5-7%), carbonate (up to 5%), hydromica, chlorite and epidote. The mineralization consists of pyrite, haematite, galena and sphalerite, as well as other minerals in quantities ≤1%. The main economic mineralization is concentrated in quartz veins. Other formations - stringer zones and metasomatic silicification - have relatively low gold contents, from 0.5 - 2.0 g/t.
The veins and veinlets consist of quartz, adularia and calcite and lesser amounts of epidote, hydro-mica, sericite, muscovite and chlorite. Ore minerals constitute less than 1% and include native gold, electrum, pyrite, sphalerite, galena, chalcopyrite, pyrargyrite, acanthite, fahlores and traces of other minerals. Gold is present as inclusions in quartz, pyrite, sphalerite, galena, and chalcopyrite. Grain size is from tens of microns to 1 mm, rarely larger. The distribution of gold and silver in the ore bodies is extremely uneven; the gold content reaches 3,300 g/t and silver 16,300 g/t; coefficients of variation for individual ore bodies, respectively, reach 376 and 632%. The Dvoynoye mine belongs to the gold-silver type (the gold:silver ratio is 1:2) of shallow gold formation according to the classification of Petrovskaya (1973). The reserves of the Dvoynoye mine amount to 64 t of gold, 94 t of silver.

The mine has been developed commercially since 1996 as an open pit. Ore zone 1 is currently mined out for the most part. As a result of exploratory work in 1995-2005, economic reserves were approved for zones 8 and 37. In 2009, exploration work was completed at Dvoynoye mine. These investigations showed that ore zones 37 and 38 are a single structure and merge into ore zone 37. Exploration of zone 37 resulted in the following amounts of gold in reserves: in category C1 - 22 t and in C2 - 43 t. The mine was re-opened as an underground mine by Kinross Gold Corporation on October 10, 2013.

Nezhdaninskoye gold mine

The Nezhdaninskoye gold mine is located 160 km from the Tomponsky District centre, near Khandyga village, in the Tyra River valley. It was discovered in 1951 by the Dybinsky Geological Prospecting Party led by G.F. Gurin. The total area of the deposit is 60 km²; it is one of the largest Russian gold prospects in terms of refractory ore.

The ore field is located within the southern part of the South Verkhoyansk synclinorium and has a two-tier structure. The lower tier is composed of the Verkhoyansk complex of terrigenous rocks (Upper Carboniferous - Middle Jurassic) up to 11 km thick, while the upper tier consists of Upper Jurassic-Cretaceous volcanic strata. The mine is on the northern flank of the Allakh-Yun gold-bearing band at the junction of the Western...
and Central structural zones of the synclinorium. It is confined to the crest of the Dybinsky anti-
cline at its intersection with major N-S, E-W and NE-striking faults (the Kiderkinsky, Tyirinsky and Suntarsky faults). The mineralization is lo-
cated in Lower-Upper Permian silt strata (Figure 13), regionally metamorphosed to lowermost
greenschist facies. There are two formations: the Dzhuntaginsky Formation which is 600-700 m
thick and is overlain by the Dybinsky Formation, 1,400 m thick. The Dzhuntaginsky formation
hosts the ore, and is underlain by shale grading into sandy siltstone with sandstone beds. The
Dybinsky Formation comprises carbonaceous
siltstone and argillaceous sandstone with silt-
stone beds. Igneous rocks are uncommon: they
include the Kurumsky granodiorite massif to the
northeast, a dyke complex, diorite porphyry and

Figure 13: Geological
structure and sections
of the Nezhdaninsky
mine area (Konstanti-
1: U. Permian/triassic
siltstone-sandstone
sediments (P\textsubscript{2}+T); 2:
U. Permian sandstone
formation (P\textsubscript{2}); 3: U.
Permian siltstone for-
mation (P\textsubscript{1}); 4:
L. Permian shale formation (P\textsubscript{1}); 5:
Marker sandstone bed;
6: Gabbrodiorite stock;
7: Dykes of intermediate
composition; 8: Diagonal
fault; 9: Nezhdaninsky
system breaks; 10:
Poperechnaya system
breaks and fracture
zones; 12: Geological
boundaries.
lamprophyre. The crest of the Dybinsky anticline is strongly deformed and broken up by numerous discontinuous displacements, corresponding to the deep regional faults mentioned above.

The main orebearing structures are splays of the Kiderkinsky (Main) fault which are trending N-S and dipping westwards at 65°. Two main morphological types of ore are distinguished in the mine: 1) extensive mineralized cataclasites composed of altered cataclastic and metamorphic rocks with dissemination and veinof sulphide mineralization; 2) quartz veins and vein zones with variegated structure. Both types of ore are accompanied by mineralized veinlet zones acquiring a joint-node stockwork appearance. In total, there are about 80 steeply oriented ore bodies, of which ten are major. The richest mineralization is localized in tabular quartz veins formed under tension, in which the ore zones form splays. Some of these can be traced down-dip for more than 550 m (vein 14). Along strike, in some cases, they split into two - three veins, which may then be transformed into a system of veinlets. The zones are 270-3,500 m long and 3.9-11 m thick; veins and vein zones are 50-340 m long and 1.3-3.4 m thick respectively. Over 90 % of the mine reserves lie in ore zone No.1 which is 3,500 m long with an average thickness of 11 m. The total vertical extent of the mineralization is 1,330 m.

The ore contains 6 % sulphides with 3 % arsenopyrite, 1 % pyrite, 0.9 % sphalerite, 0.6 % galena, 0.3 % fahlore and 0.1 % chalcopyrite and pyrrhotite. Nonmetallic minerals comprise: 53 % quartz, 29 % clay and feldspar minerals, 6.8 % mica, 3 % carbonates, and 1.7 % graphite. Only the gold and silver are of commercial value. The gold occurs in native state and is associated with sulphides and quartz. The size of the gold grains is 0.002-1.2 mm, fineness 680-840 ‰ and the ratio of gold to silver is 1:3 - 1:10. Silver occurs finely dispersed and scattered in sulphides.

Development of the Nezhdaninskoye mine began in 1974 and it was in production for 25 years. About 25 t of gold were produced over these years (from gold-quartz ore). Gold reserves in the mine as of 01.01.2014 were: proven (categories A+B+C1): 278.7 t, tentatively estimated (C2): 353.3 t, with an average gold content in the ore of 4.89 g/t. The ore is mostly refractory (arsenic).

As of 2014, the license for the mine development belonged to OJSC "South-Verkhoyansk Mining Company" (subsidiary of JSC "Polyus Gold"). Further geological exploration and a feasibility study for development of the mine are currently being implemented.

**Natalka gold mine**

The Natalka gold mine is in the Tenkin District of the Magadan Region, 450 km from the town of Magadan. The ore field has an area of 40 km² and is part of the Omchak ore cluster, the largest in the Kolyma-Chukotka metallogenic province. The field was discovered in 1942 by V.P. Mashko during a detailed survey along the Omchak River, Geologichesky and Pavlik creeks. The Omchak ore cluster is part of the Ayan-Yuryakh anticlinorium and is confined to the margin of an assumed pluton. (Goncharov V.I. et al., 2002). The formation of the ore cluster is related to the collision stage of the Yana-Kolyma fold system. Sedimentary rocks hosting the ore cluster are represented by Permain and Triassic deposits of pelagic mudstone, fine-grained sandstone, and diamictite. In places, the terrigenous sediments are enriched in syngenetec carbonaceous matter (Berger, 1990). Intrusive rocks are present at the periphery of the ore cluster, and include gabbro, leucocratic granite and dykes of various composition. To the north is the large (230 km²) Nerchin granodiorite massif.; The Mirazh and Tengkechan granite massifs as well as small gabbro and diorite bodies and dykes of various composition are located on the west and south of the cluster.

The Natalka mine area comprises Late Permian sedimentary rocks, which is divided into three formations (Figure 14). The lower Pioneer Formation is represented by carbonaceous-argillaceous shale with interbedded calcareous sandstone and gravelstone, the middle Atkan Formation is composed of tuffaceous diamictite and shale (Goncharov, V.I. et al., 2002). The upper Omchak Formation is composed of tuffaceous diamictite and shale (Goncharov, V.I. et al., 2002). The upper Omchak Formation is, in its lower part, represented by siltstone, shale, and gravelstone, in the upper part by sandstone and silty-clayey shale. All the sedimentary rocks of the Natalka ore field, regardless of age, were subject to regional greenschist facies metamorphism; secondary alteration is expressed by newly formed chlorite, sericite, pyrite, rarely stilpnomelane.
Figure 14: Geology of the Natalka mine (Goncharov V.I. et al., 2002).
Sedimentary rocks of the ore field are folded, the Natalka syncline being the major fold structure. This is a second order fold (with respect to the Tenkin anticline (Goryachev N.A. 1998)), with a simple symmetrical shape; its limbs are dipping 40-50°, the axis is trending NW, extending 4.5 km with a width of 2.5 km. The core is composed of the Omchak Formation rocks which outcrop in the central part of the mine.

The fault tectonics at the Natalka mine largely determine its structural plan. Faults which are longitudinal with respect to the fold structure or transverse to it predominate. The main tectonic unit of the ore field is the Natalka fractured zone, consisting of several faults, the most important being the Glavny and Severo-Vostochny faults, and splays of the Omchak fault accompanied by a large number of smaller joints (Goncharov V.I. et al, 2002).

The gold mineralization is controlled by these faults which affect the southwestern limb of the Natalka syncline. The ore zones are characterized by a predominance of sulphide-disseminated and linear-stockwork mineralization. The linear stockworks are represented by two systems (longitudinal and diagonal) of veinlets up to 3 cm thick, grouped around the tectonic joints in the axial parts of ore zones. The maximum thickness of the veins reaches 10-20 cm; their number/meter ranges from 5-10 to 40-50 (Goncharov V.I. 2002). Along strike and down-dip veinlets commonly grade into short, lenticular or parallel selvage veins, rarely, into zones of silification. The ore zones are subdivided into two branches forming, in a vertical cross-section, a fan-shaped system with a tendency to converge at depth.

The mineralization is unevenly distributed in four ore elongate bodies (Kalinin et al., 1992) of different shape and size (Figure 15). The richest mineralizations are found in the volcanic-sedimentary rocks of the Atkan Formation and carbonaceous-clay shales (Corg up to 2.44 %) of the Omchak Formation. In general, the gold content remains constant with ore body thickness increasing with depth (Kalinin et al., 1992). The main ore type is sulphide-dissemination in a host rock of variably altered terrigenous rocks (Goncharov, 2002). Wallrock aureoles are closely linked to alteration, including silification, sericitization, carbonatization and chloritization. Gold is consistently found in arsenopyrite and pyrite, in quantities exceeding tens or hundreds of g/t, but it is very rarely observed microscopically. The disseminated sulphides contain, according to calculations by R.A. Eremin and A.P. Osipov, 50-70 % of the total gold content.

The gold is coarser in quartz veins and veinlets; it is represented by cloddy, veinlet-like, tabular,
spongy, dendritic, and crystalline forms up to 2-3 mm in size. Two gold generations are identified: an early generation with a fineness (ppt) of 700-800 units, closely associated with pyrite and arsenopyrite, and a late generation (600-700 units), associated with galena, chalcopyrite, and sphalerite (Goncharov, 2002). Other minerals include pyrrhotite, albite, adularia, ankerite, dolomite, and quartz. Scheelite is observed occasionally. Fahlore (2-12.7 % Ag), bournonite, boulangerite, antimonite, millerite, cobaltite, loellingite are very rare.

Potassium-Argon dating of post-ore and pre-ore dykes (115-55 Ma and 155-130 Ma respectively) limits the age of the mineralization to not younger than the Early Cretaceous (Goncharov V.I., 2002).

Ore extraction and gold production at the Natalka mine started in 1945; the licensee was the Matrosov Mine. Currently, the license for geological exploration and gold production at the Natalka mine belongs to JSC "Polyus Gold", which is constructing a mine and gold recovery plant. According to investigations in 2006, the reserves within the pit outline for categories B+C1+C2 amount to 1,449.5 t of gold with an average grade of 1.7 g/t in the Natalka mine.

**Dukat silver mine**

The Dukat mine is located in the Omsukchan district of the Magadan Region, 31 km from the administrative centre of the district and 595 km from Magadan. The prospect was discovered in 1968 by geologists of the Omsukchan Expedition. The silver reserves are the largest in Russia and among the largest in the world. The Dukat ore field is located within the Okhotsk-Chukotka volcanic belt in the Balygychan-Sugoy volcanic depression, the basement of which is composed of intensely dislocated sediments of the Triassic-Jurassic Verkhoyansk complex. The depression is filled by weakly deformed volcanic and sedimentary rocks.

The ore field is situated in a volcanic-intrusive dome complex, 35 km² in area, mainly composed of Early Cretaceous felsic volcanic rocks (Konstantinov, 2003). The dome includes a large sub-volcanic body and smaller, sub-volcanic and extrusive bodies (Konstantinov, 2003). The rocks of the Dukat ore field consist of Late Triassic grey marine terrigenous molasse sediments, an Early Cretaceous mineralized rhyolite association and coal-bearing molasse formation, Early-Late Cretaceous andesite and diorite-granodiorite associations, Late Cretaceous rhyolite and leucogranite associations and a Palaeogene basaltic association (Figure 16). The central part of the ore field is composed of ultrapotassic rhyolite, ignimbrite and their tufts with horizons of black mudstone (Askoldinsky Complex). Along the periphery, there are coal-bearing deposits of the Omsukchan Group, unconformably overlain by a gently dipping andesite cover with conglomerate, tuff and rhyolite horizons. Subvolcanic bodies of various shapes, fissure intrusions and rhyolite dykes, automagmatic breccias of rhyolite, aphyric and nevadite rhyolite, diorite stocks are rather common to the north of the dome structure, forming a semicircle. Semi-circular rhyolite dykes are typical to the east and south. Drill holes in the central part of the ore field penetrated, at a depth of 1200-1500 m, a large intrusion of biotite leucogranite and granodiorite that has a metamorphic imprint on the Askoldinsky volcanic rocks. The uplift is dissected by series (bands) of Palaeogene basalt dykes with a NE strike. They intersect the ore bodies, indicating their post-ore age.

The main mineralized complex is composed of Lower Cretaceous ultrapotassic rhyolite and rhyolite ignimbrite with mudstone horizons. The lower Cretaceous subvolcanic rhyolite hosts chimney deposits and veins with silver and silver-polymetallic mineralization. The latter occur along the periphery of the ore field.

The Dukat ore field is dissected by a series of northeasterly- and, to a lesser extent, northwesterly-trending faults forming step-like breaks, mostly steeply dipping (70°). Ore-controlling faults divide the ore field into several blocks with different ore potential. The blocks can be subdivided into steeply dipping (60-90°) and gently dipping. The latter is less abundant (20 % of the total) and usually formed by splays of faults with larger displacement. Most of the faults have displacements of a few tens of metres, the largest displacement found being 270 m.

The ore bodies are located in pre-ore altered rocks, mainly of chlorite-hydromica-quartz composition. Their emplacement is controlled
by systems of faults. Two types of ore have been identified: mineralized zones and veins. The mineralized zones are complex formations controlled by the largest faults. They are represented by one or more stem-like veins with a breccia structure, the fragments being of mineralized pre-ore injection breccias and veinlet-disseminated mineralization in the host rocks. The mineralized veins consist of thick (3-5 m) stem-like quartz-rhodonite veins with veinlet-disseminated mineralization in the wall rock (1-3 m). The veins are confined to tectonic zones striking N-S and NE. The contact zones are sharp, curvaceous and 1-2 m thick.
The different ore bodies are characterized by pinches, bulges, cleavage zones, branching and upwards increase in thickness. They may be banded, brecciated, or massive, homogeneous. Zoning has been identified in the emplacement of mineral associations. Stage zoning with mineral assemblages of different ages is confined to different ore-localizing structures. Mineralizations of the quartz-sulphide stage are confined to structures with a northeasterly strike; those of the quartz-chlorite-adularia stage to structures with a near N-S strike, and rocks of the quartz-rhodonite stage to structures with a northwesterly strike. Mineral associations of an earlier gold-silver stage are typical of upper horizons of the prospect: those of the later stages are present in the lower horizons.

There is a clear lithological control on the distribution of the mineralizations. The greatest amount of ore is concentrated in the lower rhyolite, ignimbrite and aphyric rhyolite. The mineralization is rich but uneven. The great heterogeneity with respect to the structures and composition of the rock sequences has caused multi-stage screening of the gold-silver mineralization. The geological and structural position of the ore bodies is determined by various factors including: joints in multidirectional mineralized structures, areas of intersection with large ore-controlling faults, bends in mineralized faults, and intersection of pre-ore fractures filled with basaltic dykes during the post-ore stage.

There are three main ore associations: quartz-sulphide (1 %), quartz-chlorite-adularia (50 %), and quartz-rhodonite (49 %) associations. Almost all the mineral associations are separated by distinct tectonic features. The Au/Ag ratio in the quartz-chlorite-adularia type is 1:340, and in the quartz-rhodonite type, 1:550. More than one hundred minerals of hypogene and supergene origin are known in the deposit. Acanthite, native silver, kustelite, electrum and native gold are of particular economic importance. Galena, silver sulphosalts, sphalerite, chalcopyrite, magnetite and others occur in subordinate amounts.

The most common structures in the ore are banded, rhythmically banded, encrusted, scalloped, colloform and brecciated. So-called “leopard” veins (Konstantinov et al., 2003) are typical of quartz-chlorite-sulphide ore bodies, formed by crushing and later cemented with quartz-chlorite-adularia material.

Rb/Sr geochronology (Konstantinov, 2003, Sidorov, 1985) was carried out on magmatic rocks in the mine and the age of the Askoldinsky rhyolite complex was found to be 123.5 Ma, that of the granites is 80-86 Ma, and the 87Sr/86Sr ratio is 0.7119 (+/-) 0.0014, which is typical of granites of crustal origin. The radiometric ages of the quartz-chlorite-adularia and quartz-rhodonite types of mineralization are 84 and 74 Ma, respectively.

Development of the Dukat prospect coincided with the commissioning of the Omsukchan gold recovery plant in 1980. In 1995, the work on the field was suspended. 2,500 tonnes of silver were produced from 1980 to 1995. In 2002, the Close Joint-Stock Company “Magadan Serebro” (a subsidiary of “Polymetal”) obtained a license for the right to exploit the Dukat deposit and reopened the mine.

Kyuchus gold deposit

The Kyuchus mine is located in the northern part of the Verkhoyansk Region, in the Republic of Sakha (Yakutia), N of the Arctic Circle, in the lower reaches of the R. Kyuchus, a tributary to the R. Yana. The Kyuchus ore field was discovered in 1963 by the Central Survey Expedition of the Yakut Territorial Geological Department in the course of geological surveying at 1:200,000 scale.

The ore field is located on the eastern limb of the Central Kular anticline cut by the major Yana fault which separates two major tectonic elements of the Verkhoyansk-Chukotka area of folding – the Kular fold-block uplift and the Polousnensky Synclinorium. The ore field is delimited to the northwest by a low-amplitude northeasterly-striking thrust, and to the southeast by a steeply dipping normal fault parallel to the Yana fault zone, and to the southwest by a sub-latitudinal fault that separates it from the block of gentle folding (Figure 17). Two systems of northeasterly-striking faults intersect at an acute angle and dipping towards northwest at an angle of 60-80° are the main elements of the ore field structure (Moskvitin S.G., 2002).
The rocks of the ore field are represented by alternating siltstones and mudstones of Triassic age, characterized by high clay contents and carbonaceous matter. The lower member consists of siltstone interbedded with mudstone and a layer of fine-grained polymictic sandstone. The unit is finely lamellar with inclusions of marcasite and carbonate-siltstone concretions. The upper member consists of siltstone interbedded with numerous layers of light to dark grey sandstone. The sedimentary rocks in the ore field are folded in a north-south trending, anticlinal fold (Konishev V.O., 1995).

The morphology of the ore bodies is compatible with the lithological composition of the sedimentary rocks. Thin, branching shear zones are present in the northeastern part of the field, affecting sandy sediments of the upper member. The ore bodies are dispersed in the sandstone units as thin lens-like deposits. In the middle part of the field, the mineralized zone intersects sediments of the lower member, in which thick persistent shear zones (most favourable for gold-sulphide mineralization) have formed. The morphology of the ore bodies is tabular, commonly lens-like, as is inherent throughout the whole interval of crushed rocks.

Figure 17: Model of the Kyuchus ore field (Konishev V.O., 1995). 1: Ladinian (Upper Middle Triassic) rocks: a: on plan; b: in the cross-section; 2: Faults; 3: Mineralized zones of cleavage, brecciation, and boudinage; 4: Tectonic sutures of the transform fault zone; 5: thick cleavage zones transverse to the transform fault; 6: Serpentinized ultrabasic protrusion.
Within the ore field, intrusive rocks are represented by a diorite dyke and porphyric dolerite. The age of the diorite dyke is $137 \pm 5$ Ma, while the age of the granitoids of the Kular batholith is $156-133$ Ma (Moskvitin S.G., 2002).

The ore bodies (a few metres to a few tens of metres thick) with gold-sulphide veinlet-disseminated mineralization partially or completely occupy the shear zones. In some areas, the ore deposits are often disintegrated into series of branching ribbons.

Quartz-stibnite veins are controlled by areas of brecciation and mylonitization of the mineralized zones. They occupy a pivotal part of the shear zones. In plan, the mineralized veins are lenticular and lensoid, with sharp contacts. They are characterized by bulges and twists. The vein thickness ranges from 0.1 to 0.5 m and reaches 1-2 m in the bulges. There are ore shoots (enriched in gold, antimony, and mercury) in gold-sulphide ore bodies. In the longitudinal plane of Ore Body 1, the ore shoots occur as ribbon-like strips inclined northeastwards.

There are three morphogenetic groups of hypogene ores: mineralized shear zones, dilational veins, and breccias. The gold-bearing mineralized shear zones include disseminated and veinlet-disseminated sulphide ores. Disseminated sulphide (1-5 % sulphides) ore is represented by hydrothermally altered sandy siltstone and mudstone with fine, dusty dissemination of needle crystals of arsenopyrite and cubic pyrite. This type of ore makes up the outer zone of the ore-metasomatic system. Veinlet-disseminated sulphide ores fill the inner part of mineralized shear zones and are the main economic ore type in the mine. Newly formed vein minerals are quartz, carbonate, and sericite. A distinctive feature of the ores is the combination of a high frequency of quartz-carbonate veinlets and abundant sulphide dissemination in altered, crushed rocks. Breciated sulphide-stibnite ores make up selvages of quartz-stibnite veins and areas of intensive brecciation of sulphide mineralization zones. They control the superposition of complex gold-stibnite-cinnabar mineralization on ores of the earlier gold-sulphide stage.

The ore mineralization consists mainly of pyrite, arsenopyrite, stibnite, cinnabar, and rare native gold and mercury. Finely dispersed gold is concentrated in pyrite and arsenopyrite; the economic grade in these ores reaches 100 and 440 ppm respectively. The concentration of dispersed gold in stibnite is never more than 5 ppm (0.5 ppm on the average). From the ratio of gold and silver concentrations in the minerals, the estimated purity of the gold is 800-850 ‰. The inferred gold reserves in the Kyuchus mine are 178 t with an average grade of 8.5 ppm; probable resources are 120 t. By-product grades are: 1.5 ppm Ag, 1.7 % As, 0.5 % Sn, 0.024 % Hg. The prospect can be developed both in open pits and by underground mining. In 2012, the prospect, with a total area of 225.5 km², including the Kyuchus mine, was put up to tender, but the tender was not held because of lack of applications for participation. Currently, the ore field belongs to the state.

**Prognoz silver deposit**

The Prognoz deposit is located north-east of Yakutia, within the Yana-Adycha ore district in the transpolar part of the Verkhoyansk Range on the divide of the Sartang and Nelgese Rivers, tributaries of the R. Adycha. The deposit belongs to the Ulan-Chaidakh ore cluster. It was discovered by Yu. N. Badarkhanov and was first prospected with trenches by V. S. Prokopiev and B. N. Podyachev (1973-1977.)

The deposit is located within the Verkhoyan-Kolyma fold system in the eastern part of the Sartang synclinorium near its junction with the Adycha brachyanticline. Sets of major fractures have near N-S and near E-W strike directions. The former are concordant with the folding, occur in interstratal strips, and separate tectonic blocks with linear folding and blocks with horizontally bedded rocks. The fractures of near E-W and north-easterly strikes, with which the mineralization is related, are later features.

The Prognoz ore field is hosted by Middle and Upper Triassic sandstones and siltstones that are overlain by unconsolidated Quaternary sediments 2-10 m thick (Figure 18). The deposit is confined to the arch of a north-trending anticline. Intrusions within the ore field (Cretaceous quartz porphyry and lamprophyre dikes) are subordinate. The ore bodies show cross-cutting relationships to their host rocks;
they are represented by mineralized cataclasite zones with no clear boundaries. The central parts of the ore bodies consist of breccias, brecciated sandstones, quartz-carbonate-sulphide (sulphosalts) veins, and the peripheral parts of cleaved rocks with quartz-carbonate-sulphide veinlets.

The major part of the deposit is represented by the Glavnoye ore body, which contains over 50% of the reserves and probable resources of silver. Its ascertained length is 4000 m, the maximum commercial mineralization depth is 250-270 m and its thickness is up to 18 m (average: 4 m). It has a near E-W strike (90-120°) and steep southward dip (70-80° up to vertical). Silver is distributed unevenly, with the largest concentrations confined to ore shoots whose positions are controlled by fractures across the body. The content of silver in the ores varies from 10 g/t to 28 kg/t, lead from <10 % to 52 %, and zinc from <10 % to 15 %. The central part of the Glavnoye ore body contains a field of the richest ores, which was given its own name: Izgib.

The Boloto ore body is 500-800 m S of Glavnoye. It is the deposit’s second-most productive ore body and includes 21 % of the reserves and probable resources of silver. It has been traced along strike for 2300 m, and for up to 260 m down dip, with thicknesses of up to 7 m (average: 2 m). The content of Ag in the ores is up to 17.7 kg/t, of Pb up to 25 %, and of Zn up to 21.5 %. Apart from Glavnoye and Boloto, nine other ore bodies with NE and near E-W strikes have been discovered in the deposit. Their parameters are: length: 300-650 m, thickness: 1-4 m, average content of silver: 370-1170 g/t, lead: 0.1-5.0 %, and zinc: 0.1-2.0 %.
The ores of the Prognoz deposit are silver-polymetallic and moderately sulphidic. Their principal minerals are carbonates, quartz, galena (in particular, silver-bearing), sphalerite, Pb-sulphosalts (in particular, silver-bearing). Silver minerals include miargyrite, freibergite, ruby silver, owyheeite, diaphorite, schirmerite, zoubekite, argentite, and native silver.

The deposit development license was owned, in 2014, by Prognoz-Serebro LLC. The silver reserves were, as of 01.01.2014: proven (Category A+B+C1), 4224.5 t, with an average silver content in ores of 906.3 g/t; inferred (C2), 4966.0 t. Prognoz was, as of 01.01.2014, Russia’s largest true silver deposit in terms of silver reserves (larger than Dukat). In Russia, more silver is concentrated only in the silver-bearing Udokan giant copper deposit (No. 1 in in-situ reserves). The average silver grade in the Udokan ores however is only just over 1 % of that in the Prognoz ores, and silver can be mined at Udokan only as a by-product from copper mining.

Blagodatnoye gold deposit

The Blagodatnoye deposit is located in the Central Sikhote-Alin (Khabarovsk Krai). The ore field is confined to a cryptobatholithic zone of the early phase of the Prisikovsky granitoid massif consisting of monzogranodiorite and granodiorite with a high magnetic response, mostly of a sodic type, with high iron oxidation factors (0.33–0.6), and intruded into a series of Lower Cretaceous sandy/clayey rocks. These rocks are, in addition, intruded by dioritic stocks and by many diorite porphyry dikes. The rare-metal mineralization (Sn, W, Mo, Be) is closely related to granites of the second phase (non-magnetic, potassium) which is of late Cretaceous age. The deposit consists of multiple mineralized cataclasite zones containing rare thin quartz veinlets, and veins of quartz, quartz-carbonate, and quartz-sulphide. The mineralized zones’ thickness is 0.1-0.9 to 3.6 m. Fragments of sandy/clayey rocks in the zones are cemented by quartz, carbonate and limonite; siderite, and pyrite and arsenopyrite impregnation are observed. The ore veins are complex, branching bodies with many apophyses and pinches. The vein thickness is 0.1-0.3 m (up to 0.5-0.6 m). They are dominated, near the surface, by quartz; apart from quartz, carbonate occurs in the underground mines. Sulphides, including pyrite, arsenopyrite, sphalerite, galena and, less commonly chalcopyrite and molybdenite make up 3-5 %. Impregnation (up to 0.5-1 mm across) of straw-yellow gold is observed. The length of the main mass of veins is a few tens of metres, locally up to 100-200 m and rarely up to 500 m. Their strike may be north-easterly or near north-south, less frequently north-westerly and their dip 50-90°. The average gold contents in the deposit are 1.5-2 g/t. Ore shoots stand out at vein intersections with mineralized zones of north-easterly and north-westerly strikes. The deposit has been partially mined.

The license for prospecting and mining in the Blagodatnoye deposit is owned by Highland Gold Mining Limited (Russdragmet). A multifaceted exploration program is intended to verify the proven reserves. The deposit’s forecast resources are 5 tons of gold.

Kupol gold-silver deposit

The Kupol deposit is at the border of the Bilibino and Anadyr Districts of Chukotka Autonomous Okrug (AO) in the north-west of the Anadyr Highlands, near the R. Keyemraveem (tributary of the R. Mechkereva), 300 km from Bilibino. The deposit was discovered by V. V. Zagoskin in 1995 during a 1:200,000 scale geochemical survey. Supplementary exploration was carried out in 1996 and detailed prospecting from 2003.

The deposit is in the northern part of the Mechkereva depression of the Okhotsk-Chukotka volcanic belt deformed by younger tectonic subsidence. It is confined to the the Middle Kayemraveem N-striking fault zone. The deposit occurs in an area of Upper Cretaceous andesites, andesite-basalts, and their tuffs. Intrusive bodies in the district include diorite-porphyry plutons and felsic to mafic dikes, including rhyolites, microgranites, andesites, andesite-basalts, and micro-gabbros (Figure 19). The thickness of the penetrative Middle Kayemraveem fault zone in the ore field is about 150 m, and it dips 75°E.

2 This deposit should not be confused with the Blagodatnoe deposit in Krasnoyarsk Krai.
The enclosing diorite-porphyry, andesites and granodiorite porphyries in the western footwall of the fault are intensely altered by hydrothermal and metasomatic processes. The rocks are well-developed quartz-sericite metasomatites and contain epithermal ore-bearing veins of quartz and, less frequently, of carbonate-quartz, with weak sulphide impregnation (less than 0.5%). The ore-bearing vein zone is over 3.1 km long, its thickness is 10-30 m, and the length of the area with ascertained economic mineralization is about 1000 m. The mineralization has a vertical extent of over 430 m.

Sixteen ore bodies have been identified in the deposit. They are represented by quartz veins and, to a much lesser extent, by breccias with quartz cement. The veins are not persistent in strike and dip, and are partly replaced by breccias or veinlet silicification zones. The strike of the veins is N-S, concordant with the regional tectonic strike; the dip is 75-85°. The ore bodies are 2-32 m thick and 50-2300 m long. The gold and silver distribution in the ore bodies is extremely uneven. The gold content in the ores varies from 0.01 to 100.0 g/t (up to 2622.1 g/t), the average being 21.5 g/t. The silver content is 0.5-10.0 to 500 g/t (up to 32417.3 g/t), the average is 266.6 g/t. The highest gold and silver contents are confined to quartz and carbonate-quartz breccias. The gold occurs in the ores as nuggets. The gold-silver ratio in the deposit varies from 1:1.6 to 1:50, with 1:10 - 1:11 on average.

The gold-silver mineralization of the Kupol deposit occurs in hydromica-adularia-quartz veins and vein-veinlet zones. The vein minerals are mainly quartz (up to 75 %), sericite, white agate, chlorite, kaolinite, montmorillonite, and zeolites. Ore textures include colloform-banded, concentrically banded, scallop-banded, and sheaf-banded breccia combined with framework-platy, cockade, crested and drusoid textures. The latter is typical for amethyst and crystal quartz. The ore aggregates are mainly thin/fine-grained, less frequently medium-grained. The alternation of bands and ore mineralization distribution in them is determined by the varying textural/structural features of ore aggregate structures, with ovoid, globular, and columnar-crested unevenly recrystallized texture. A feature of the ores is their low content of ore minerals, 1 % or less, though in certain vein sections their content reaches 3 %. Apart from native gold, various sulphides occur in the ores: pyrite, pyrrhotite, chalcopyrite, sphalerite, arsenic and antimony sulphosalts of silver and copper, less frequently gold- and silver selenides. The main economic ore minerals are native gold (mainly 610-700 fineness), ruby ore, stephanite, tennantite, freibergite, acanthite and pearceite. Naumannite and aguilarite occur, but are rare.
The gold occurs as nuggets, varying in size from <0.01 mm to 2.2 mm; the main bulk of the gold in the ore (75 %) is found in fine (less than 0.074 mm) fractions. Native gold occurs in aggregates with vein- and the ore minerals as listed above. The reserves at Kupol, after a geological survey by ChGGP JSC in 2004-2005, were found to be: gold: 154.8 t, with an average content of 18-20 g/t, and silver: 1875 t, with average contents of 220-260 g/t, which make it a large deposit. The mining and gold recovery plant were opened in 2008. The plant’s daily capacity is 3,000 t of ore. As of 2011, 45 % of the deposit’s active reserves had been mined.

Pavlik gold deposit

The Pavlik gold deposit is located in the Tenkinsky District of Magadan Oblast, by the Omchak stream where it joins the R. Tenka, 20 km S of the Natalkino gold field. The site was discovered by geologist Ye. P. Mashko in 1942: detailed exploration began in 1944, continuing until 1954. The deposit is part of the Omchak placer ore cluster (Figure 20), located in the Ayan-Uryakh anticlinorium, within the contact zone of a granitoid pluton cut by the major Tenka fault (Goncharov et al., 2002) and related to the collisional stage of the Yana-Kolyma fold system.

The rocks enclosing the Pavlik ore field are Late Permian sediments, mostly tuffaceous, silty/pelitic and clayey schists, as well as polymict and arkosic sandstones and fine-pebble conglomerates. The ore occurrences dip steeply or very steeply, complicated by minor folds and fractures. Two minor intrusions are localized on the flanks of the Pavlik ore field: the Vilkin diorite stock at the Vilkin creek, and the Vanin stock of quartz diorite, granodiorite, quartz porphyry, and rhyolite breccias between the Vanin and Lyotchik creeks. Several dykes of intermediate composition (spessartites, porphyrites) are found, and more rarely acidic rhyolites and granite porphyries. The dykes strike NW and dip SE and NE at angles of 50-80°, and are 0.5-1.5 m thick.

The fracture tectonics in the deposit area are very intense, with faults varying in extent, strike, intensity and mode. Contiguous fault systems build up cataclasite zones; their thickness varies from fractions of a metre to scores of metres. The internal ore tectonics are defined by the wide development of local shear zones which typically, in the postore stage, show reactivation with amplitudes of 0.1-20 m.

About 30 ore zones are known within the Pavlik ore field, of which two represent the bulk of the ore. The mineralized zones have a north-westerly strike and a thickness of 5-10 m to 30-40 m. The ore zones are consistent in strike and dip, but their thicknesses vary; the average thickness is 1.5 - 7.0 m and the average gold content varies from 3.4 to 6.6 g/t. The ore zones are a combination of veins, veinlets, and metasomatic-breccia-silicification zones; the mineralizations commonly form stockworks. Selvages of the ore zones are unclear and can only be identified by analytical data. The deposit has blind ore zones, some of them outcropping as veinlet systems with weak mineralization, which develop into thick ore bodies with depth. In terms of morphology, the ore bodies in the Pavlik deposit occur as:

- Veins and lenses of open-cavity filling with clear, sharp contacts.
- Systems of near-parallel branching quartz veinlets.
- Silicified intensely crushed zones.
- Belts of hydrothermally altered wallrock enclosing rocks.

The mineralogy and genetic attributes of the Pavlik deposit mineralization indicate that it belongs to the low-sulphide (pyrite-arsenopyrite) type of gold-quartz deposits. The veins are mostly of quartz, commonly quartz-carbonate and rarely only carbonate. Apart from native gold, the ore minerals observed in bulk are arsenopyrite and pyrite. The former is more frequently found in vein quartz, but is also widespread in the enclosing rocks as small scattered disseminations, and less frequently as nodular accumulations. Pyrite is more typical for the host rocks, the total sulphide content not exceeding 1 %. Gold is observed consistently with arsenopyrite; its distribution is uneven or very uneven; it occurs as irregular grains, lamellae and flakes. The particle size varies from <0.1-0.5 mm and, rarely, from 1.5-4.0 mm.

Currently, the license for geological surveying and mining of the Pavlik deposit is held by Arlan...
Investment Co. An open pit and concentrating mill are being built, for a capacity of 3 Mt of ore per year. The deposit reserves in Category C1+C2, based on company data, are estimated at 154 t of gold, with probable reserves of over 200 t.

**Ametistovoye gold deposit**

The Ametistovoye deposit is located in the northern part of the Kamchatka Peninsula, 100 km NW of Korf. The deposit was discovered in 1968 during a 1:200,000 scale geological survey. The Ametistovoye deposit is located in the Tklavayam ore field, in the western part of the Ichigin-Unneivayam volcanogenic district, a fragment of the Palaeogene West Kamchatka volcanic belt.

The district’s magmatic formations are grouped into two complexes (Khvorostov & Zaitsev, 1983): Unnei, of acidic composition and crustal origin, and Ichigin, intermediate and of deep-seated origin. The base of the section (150–200 m) is represented by rocks of the Unnei complex: ignimbrite-like liparites, liparite dacites, acidic tuffs overlain by tuffites, and tuff sandstones with carbon streaks. The top of the section, represented by the Ichigin complex, comprises lavas and tuffs of dacites, andesitic dacites, andesites and andesite basalts (200-400 m), and by subvolcanic bodies of diorite porphyrites, andesites, and dacites forming the central and peripheral orifices of the original volcano.

The Tklavayam ore field (Figure 21) is confined to a volcanic structure (of the same name) of a central type, intruded at the intersection of the near N-S trending Uchigin and north-westerly-trending Tklavayam regional faults. The ore field has a complex structure; it is split into several sectors by fractures of northwesterly and near N-S strike and blocks sunk stepwise into the central part of the structure. These fractures, together with the extrusive and subvolcanic bodies, are the basic structural elements of the ore field.

The mineralization is localized in the centre of a palaeo-volcano in subvolcanic intrusions of diorite porphyrites, dacites, andesites, trachydacites, which are propylitized and adularized. The ore bodies are parts of vein bodies and vein zones containing economic grades of gold and silver, primarily by cavity filling, with subordinate metasomatic replacement. The ore zones’ thicknesses are up to 100 m, and their length up to 1.5 km. The ore bodies’ length varies from <100-800 m, their thickness being 0.5-2 m, and 3-6 m in swells. The ore bodies’ strikes are diverse: E-W, N-S, NW and NE. The veins dip steeply, and consist predominantly of quartz with or without sulphides. The veins are grouped in zones, 4-6 veins in each. The largest veins are Champion, Izyumin-ka and Ichiginskaya. In all, about 300 veins are known, with economic ones located in an area of 5 km², which constitutes only one-seventh of the area of the ore field. The structures and composition of the vein fillings are diverse, but banded and rhythmically banded textures predominate. Four ore deposition stages have been identified: early silicification, sulphide-quartz, gold-kaolinite-sulphide-quartz (main productive stage), and carbonate-quartz. High selenium contents (30-40 g/t) are observed in the ores, and up to 1 % of total lead, zinc, and copper.

The veins consist of quartz with lesser quantities of kaolinite and chlorite; less common minerals include illite, calcite, montmorillonite, dickite, and adularia; fluorite and zeolite are present, but rare. The sulphide content is up to 3-5 %, but up to 20-30 % at depth. The ore minerals are electrum, acanthite, and more rarely, ruby, silver, polybasite, stibiopearceite, stephanite, miargyrite, proustite, plagionite, naumannite, aguarilarte, kustelite, native silver etc. Pyrite, galena, sphalerite, and chalcopyrite are also quite abundant in the ores. The Au:Ag ratio is commonly 1.5-2.1, and up to 1:10-20 in sulphide ores.

Native gold is found in quartz, in aggregates with acanthite; gold grains abut locally on silver-bearing minerals. The size of the gold particles is 0.001-0.01 mm. Isometric cloddy and flattened cementation-honeycomb forms predominate; the smallest, cub-octahedral and dendroid crystals up to 0.05 mm in size, are typical. Gold assays (fineness) vary from 200 to 680, with 550-600 predominating; at deeper horizons assays increase to 730-875, and 830-960 in oxidized ores. Impurities in the gold include: Sb, Pb, As, Sn, Cu, Mn. Gold is positively correlated with Ag, Se, and Sb, and show negative correlations with Pb, Zn, Cu, and As.
The deposit's reserves were approved by the RF State Commission for Reserves on May 24, 1995; for Category C1+C2, they are 52.5 tons of gold with an average content of 13.62 g/t, the cut-off being 3 g/t. The deposit mining license is owned by Zoloto Kamchatki JSC. Development of the deposit began in 2014 and a gold recovery plant, with a capacity of 600,000 t/a, is being built. A test using a direct-leaching method gave a gold recovery level of 97.5%. The ore will be mined by both open-pit and underground methods, which will reduce mining costs and lead to more efficient attainment of the facility’s design capacity.

**Elgi River Placer (Au)**

The Elgi River placer is located in the Yana-Kolyma gold-bearing belt, in the Upper Indigirka District. The R. Elgi is a tributary of the R. Indigirka and flows across the Elgi Plateau. The placer is part of the Taryn-Elgi gold-bearing zone, 100 km away from the river head, in a 4.5–5 km wide valley. The highest water flow rate is 191
m³/s. The valley extends along a fracture zone and along the axis of a synclinal structure. The placer extends for 2.6 km, is tens of metres wide, has a lens-like structure, and comprises several channels 100-600 m long. The direction of the enriched channels corresponds to the strike of the parent rocks. The thickness of the productive strata is generally 1-2 m, seldom 3.2-4.8 m; gold is found in the parent rocks to a depth of 2 m. The linear distribution of the gold reserves changes noticeably through the placer, with the highest values recorded in the middle.

The gold is relatively uniform in the distribution of its forms. Plates and grains predominate; tablets, dendritic crystals, dendroid forms, and rods are less frequent. Plates make up 30-95 % and have varying thickness, their numbers increasing both down the placer and laterally. In the most enriched channels the content of plates and rounding of gold grains decrease. The grain sizes are distributed as follows: fine (< 2 mm), 47.1 %; medium (2-6 mm), 38.7 %; large (over 6 mm), 14.2 %. The average gold grain size is 3.99 mm; within the placer's commercial sector, grain size variations are small. The highest contents of the 4-6 mm fraction are observed on Lines 482, 486, and 490. Large-fraction gold and nuggets are found in the middle of the placer because of local mother lodes on Line 482. The gold fineness varies from 850 to 986 ‰. The average fineness for 35 analyses is 887 ‰.

Bolshoi Taryn River Placers (Au)

The Bolshoi Taryn is a tributary of the R. Indigirka located in the Taryn-Pilsky ore cluster. The placer deposit extends along ore-bearing fractures that emerge due to destruction of quartz veins and silicification zones located directly in the placer bedrock and related to gold-arsenopyrite-pyrite and partially to the gold-sulfo-antimonite mineral types. The flood plain placer has been traced for several kilometres. Its commercial channel starts from Line 532 where a quartz vein with a certain gold grade is located. The accessory minerals are pyrite and arsenopyrite, which are also found in alluvial sands. Plate, tablet, dendroid, and irregular gold grains have been found both in the placer and in the quartz vein. Silicification zones (0.12-0.5 m thick) have also served as sources for the placer; they are found not only in its upper but also in its middle part.

An important feature of the Bolshoi Taryn placer is alteration of the mineral composition of parent gold sources across the valley profile.

Berelekh River Placer (Au)

The deposit is located in the central part of the Inyali-Debin synclinorium. Its basement consists predominantly of Middle Jurassic sandy/clayey schists, cut by multiple dikes of acidic and intermediate composition and by granitoid intrusions. Quaternary sediments form a ubiquitous cover. Middle Jurassic sediments are represented by the Meredui suite, the base of which has a rhythmic alternation of grey fine-grained quartz-feldspar sandstones, lamellar siltstones and dark-grey clayey schists. The upper part of the suite consists of thick strata of grey fine-grained sandstones with streaks of clayey schists and siltstones. Sparse streaks and lenses of gravelstone and conglomerate are typical. Quaternary sediments fill the wide (up to 3 km) valley of the Berelekh and the valleys of its tributaries, and form a cover on valley slopes and divides. The sediments are of various genetic types and facies including alluvium, deluvium, eluvium and proluvium.

The district is located in the central part of the Inyali-Debin synclinorium of the Yana-Kolyma orogen. The entire sedimentary complex is deformed by linear folds with a north-westerly strike and a subsequent fracture system. The largest folds are the Solkalyakh syncline and Berelekh and Yutichna anticlines which have wavelengths of 5-10 km and lengths of tens of kilometres. They consist of linear folds of higher orders, whose axes are oriented in the same direction as those of the larger folds. The strata on the fold limbs have steep to vertical dips; inverted folds are also found. In terms of spatial orientation relative to folding and displacement amplitude, three groups of faults corresponding to three deformation stages are identified in the district: pre-intrusive, post-intrusive, and neo-tectonic.

Geomorphologically the district is located in the piedmont of the Burganda massif and occupies the northwestern part of the Talon trough, which, in addition to various exogenetic and endogenetic factors, predetermined the nature of its geomorphic structure. The Berelekh River has an approximately north-south direction and...
valley width of 4–5 km. The cross section is asymmetric, with a steep east side and flat west side. In all, seven terrace levels have been identified, 3-5 to 200-210 m high. In addition to those seen in today’s relief, buried ancient valleys are also found.

The gold-bearing stratum occurs in the lower horizon of alluvial sediments, in the eluviun and the upper part of the parent rocks, penetrating them to depths of up to 1 m. The gold distribution is pocket-and-channel; against the background of vertical reserves of 0,1-1 g/m², 1-2 g/m², and 2-4 g/m², there are channels and pockets with vertical reserves of 4-8 g/m², rarely 8-16 g/m². The gold occurs as flattened plates, flakes, tablets, or lumps, in individual rods or grains, sometimes ore-like (in fine and very fine fractions), nuggets of up to 1.5 g occur. The gold grain surface is smooth, pitted, granular, cavernous; the shapes are well-rounded oval, extended, elongated, or irregular. Gold of large and very large fractions often occurs in aggregates with milk-white quartz. The gold color is yellow with a slight reddish tint. The grains are generally well-rounded. Apart from gold, the placer contains cassiterite, scheelite, wolframite, ilmenite, arsenopyrite, pyrite, pyrrhotite, garnet, zircon, hematite, galena, and monazite.

As of 01.01.2011 the National Register included, for the Allakh-Yun River deposit (Vidny field): Category C₁ commercial reserves of 5,089,000 m³ of sand and 3,960 kg of gold, and non-commercial reserves of 349,000 m³ of sand and 146 kg of gold.

Bolotny Creek Placer (Au)

The Bolotny creek is a tributary of the At-Uryakh creek; the placer is within the At-Uryakh ore cluster. Geologically, the deposit is related to the Inyali-Debin megasynclinorium which folds Jurassic terrigenous sediments. The placer is 1170 m long, 10-80 m wide and has an average gold content of 2.2 g/m³; and a fineness of 917 ‰. The gold distribution is pocket-and-channel; the vertical reserves are represented by the classes 0.1-1 g/m² and 1-2 g/m². The gold consists mainly of tablets and plates of irregular shapes with rugged edges, less common lumpy grains; medium rounding; gold grains with a spongy surface and quartz growths can be found; the colour is yellow or reddish-yellow. Apart from native gold, the placer contains ilmenite, magnetite, scheelite, rutile, zircon, and arsenopyrite.
Rakovsky Creek Placer (Au)

The Rakovsky creek is a tributary of the River Sluchainy; the placer is part of the Orotukan ore cluster. Geologically, the deposit is related to the Inyali-Debin megasynclinorium built up by Middle Jurassic terrigenous sediments. The relief consists of a dissected low-hill terrain with absolute elevations of 600-900 m and relative elevations above the valley bottoms of 250-400 m; the creek valley is asymmetric, with remnants of buried terraces 5-10 m high.

The placer is 1200 m long, 20-90 m wide and has an average gold content of 1.47 g/m³ and a fineness of 957 ‰. The gold distribution is pocket-and-channel; grades in successive layers are 0.1-1 g/m², 1-2 g/m², 2-4 g/m² and 4-8 g/m²; there are channels and pockets with vertical reserves of 8-16 g/m², 16-32 g/m², but rarely higher.

The gold is represented by plates, tablets, tiny lumps, dendrites, flakes, or dust with wavy, smoothed, or embayed edges, with an even, tuberous, or coarse surface; the grains are well-rounded, with poorly-rounded gold occurring in fine fractions; the colour is pinkish-yellow or reddish-yellow, locally greenish-yellow, brown or black due to an iron oxide film. Apart from native gold, the placer contains pyrite, arsenopyrite, galena and chalcopyrite.

Kongo River Placer (Au)

The Kongo River flows into the R. Kolyma from the east; the placer is located within the Sibik-Tyellah ore cluster. Geologically, the deposit is related to the Inyali-Debin megasynclinorium formed by Middle and Upper Triassic terrigenous sediments. The relief of the area is dissected low-hill terrain with moraine in the upper part of the valley: summit elevations are 800-1250 m and the relative elevations above the valley bottoms are 200-500 m. The southern tributaries of the Kongo originate from the Pravo-Obninsky granite massif where summits reach 1840 m; the valley is asymmetric, and has terraces 0.5-4 m, 5-10 m, and 20-40 m high.

The Kongo River placer is a valley and banded placer, with channel ore distribution and of alluvial genesis. Commercial reserves for open-pit mining have been explored at the outwash of the Ozerny, Sukhoi, and Agach creeks with branching to the Ozerny creek valley and on the right terrace of the 10th level; on the lower flank, the placer has been partially mined (at shallow depths); the gold-bearing stratum occurs in the bottom horizon of alluvial sediments and in the decayed top part of parent rocks, to a depth of up to 1.8 m.

The placer is 6625 m long, 20-520 m wide and has an average gold content of 7.48 g/m³ and a fineness of 859 ‰. The gold distribution is pocket-and-channel; grades in successive layers are 0.1-1 g/m², 1-2 g/m², 2-4 g/m² and 4-8 g/m²; there are channels and pockets with vertical reserves of 8-16 g/m², 16-32 g/m², but rarely higher.

The gold is represented by plates, tablets, tiny lumps, dendrites, flakes, or dust with wavy, smoothed, or embayed edges, with an even, tuberous, or coarse surface; the grains are well-rounded, with poorly-rounded gold occurring in fine fractions; the colour is pinkish-yellow or reddish-yellow, locally greenish-yellow, brown or black due to an iron oxide film. Apart from native gold, the placer contains pyrite, arsenopyrite, galena and chalcopyrite.
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Blagodatnoe

Mayskoye

Dvoynoye

Nezhdaninsky

Natalka

Kyuchus

Prognoz
Sardana Lead-Zinc Deposit

The Sardana stratiform lead-zinc deposit was discovered during a 1:200,000 scale geological survey carried out by A. I. Gorbunov in 1971. The deposit is a part of the Kurung-Dyukat ore cluster. It is located near the middle reaches of the R. Aldan, in the spurs of the Gornostakh ridge extending N-S from the Tompo River to the R. Allakh-Yun River parallel to the Sette-Daban ridge. 1:50,000 scale geological mapping was carried out in 1972–1974, and detailed prospecting at 1:10,000 scale in 1972–1977. The deposit area was covered by an aerial magnetic survey for the 1:200,000 scale, partially for 1:50,000 scale mapping, and also by a 1:200,000 scale gravity survey.

The deposit is located in the Yudomian-Mai structural/formational zone at the interface between the Siberian platform and Verkhoyansk fold system. Regionally, the Sardana deposit is confined to the Kyllakh foldblock rise of the Sette-Daban horst anticlinorium. The deposit is located in the Sardana syncline, extends 9 km N-S, and has a markedly variable crosssection shape. Its width is 3.5–4 km but contracts southwards to 1.5–2 km. Second- and third-order...
folds, commonly accompanied by fractures are widely developed in the core of the Sardana syncline. Magmatic rocks are represented by rare sills and dikes of dolerite and microgabbro of Late Proterozoic age, and very rare alkaline rocks.

Three structural stages are identifiable in the district: Middle-Upper Riphean folded basement and Vendian to Palaeozoic, and Mesozoic, forming the fold nappe of an epi-Baikal platform. Within the deposit, the stratigraphic base is represented by Upper Riphean sediments (argillites, siltstones and sandstones). Terrigenous carbonate and carbonate sediments of the Vendian Yudomian suite (silica dolomites, dolomites, sandstones with siltstone streaks, dolomites with argillite streaks, limestones) overlie these. The uppermost sequence in the section consists of Lower and Middle Cambrian carbonate and terrigenous carbonate formations (dolomites, limestones, bituminous dolomites, argillites, carbonate breccias, marls). The orefield rocks are deformed into a series of linear folds with an approximately E-W strike and a marked asymmetric structure. The anticlines mostly have flat eastern limbs and steep, in some cases west-vergent limbs.

Most of the ore bodies are confined to steep or vergent fold limbs. The ore zones are both concordant and occasionally flat and with cross-cutting relationships to the lithological bedding. They have a N-S elongation and are closely associated with epigenetic dolomites. The ores are localized in hydrothermally/metamorphically altered carbonate rocks and dolomites (Figure 22). They consist of strips of impregnated and nested-impregnated sulphide mineralization with alternating poor and rich impregnation zones. Mineralization is widespread in the rocks of the upper subsuite of the Sardana suite. The ore-enclosing host unit is 80-120 m thick. There are three main groups of ore bodies in the host rock: one near the base, one in the central part and one in the upper horizons.

The lower ore level is confined to the bottom of the ore-bearing unit, and locally enters the bituminous horizon. The metal accumulation at this ore level is the largest, and is represented by galena-sphalerite and pyrite-galena-sphalerite (pyrite-polymetallic) ores. The ribbon-like banks are up to 3 m thick and up to 1500 m long. The middle ore level comprises ore bodies consisting of galena-sphalerite ores, locally with pyrite. Most of the ore bodies are confined to this level. The ore bodies form “tabular” banks about 5 m thick and 150 to 200 m long. The uppermost ore level is localized under variegated limestones, though the mineralization occasionally extends even higher. The ore bodies consist of galena and sphalerite and are usually confined to a productive horizon of dolomites, and have a limited lateral extent. The ore bodies consist of ore shoots with an average thickness of 10 m and length of 75-100 m, and are rich lead-zinc ore.

The main economic type of ore in the deposit is galena-sphalerite ore with sphalerite predominant (Figure 23). Locally there are monominerallic sphalerite and sphalerite-galena-pyrite ores. Textural features observed in the ores include massive, banded, veinlet, veinlet-nested-impregnated, veinlet-impregnated, and impregnated ores; breccia-like ores also occur, but are rare. Massive sphalerite-galena ores do not form separate ore bodies and have limited, local development. The ores are generally banded, locally breccia-textured and impregnated types occur.

The deposit’s main minerals are sphalerite, galena, dolomite, and calcite; subordinate minerals are pyrite, marcasite, sulphosalts of antimony and arsenic, arsenopyrite, and chalcopyrite. The lead-zinc ratio is 1:4 - 1:5. Cleiophane (a pale-coloured variety of sphalerite), rare jordanite, geocrontite, guadalcazarite, electrum and other minerals have also been observed. Secondary minerals developed in the oxidation zone (in small amounts) include cerussite, smithsonite,
hemimorphite, scorodite, crocoite, anglesite (lead sulphate), goethite, and hydrogoethite. The deposit is characterized by rich galena-sphalerite ores with high contents of accessory components, including germanium (20-1150 g/t), of cadmium (10 - 2900 g/t), and mercury (0.001-100 g/t). The host minerals are sphalerite for germanium, cadmium and mercury, and galena for silver. The deposit’s reserves (Category C2) were, as of 01.01.14: lead: 592,000 t, with an average grade of 3.23 %; zinc: 1,926,400 t (10.50 %). The deposit development license was in 2014 held by Siberian Nonferrous Metals LLC (a member of the Summa Group). Under the terms of the license, the target date for beginning the development of the Sardana deposit is October 2017.

Pavlov Lead-Zinc Deposit

The Pavlov deposit is located on the NW coast of South Island, Novaya Zemlya. The deposit was discovered in 1990-1995 during 1:50,000 scale geological mapping by the West Arctic crew of the Polar Marine Geosurvey Expedition. The Pavlov ore field is within the Pai-Khoi-Novaya Zemlya fold system formed in the early phases (late Triassic) of Cimmerian tectogenesis. The host rocks are marine sediments of varying composition, with subordinate igneous and igneous-sedimentary formations. Silurian and Devonian terrigenous sediments and carbonates are found in the area of the deposit, which covers >12 km². These units are folded into the large Bezmyanskaya anticline the amplitude of which is 3-4 km; the fold limbs plunge at 25º-45º S/SE. A few occurrences of magmatic rocks are found in the region of the Pavlov deposit. Sills of dolerite and gabbro-dolerite of the late Devonian Kostin Shar complex are located on the SE flank of the Bezmyanskaya anticline, outside the ore field. Mesozoic sub-alkaline picrite-dolerite dikes are also found. Both sills and steeply-dipping dikes intersect the ore body.

The Pavlov deposit is located in the area of the periclinal closure of a brachyform anticlinal fold, complicated by longlived fracture zones, primarily local thrusts with significant (a few tens of metres) displacement. The eastern limb of the Bezmyanskaya anticline is complicated by a horst structure, with a north-westerly strike, forming the deposit’s eastern block. Systems of arcuate thrusts and thrusts with north-westerly strikes are present in the south-eastern part of the ore body. Three ore-bearing blocks, the Central, Western, and Eastern blocks, are identifiable within the Pavlov ore field. The ores are confined to carbonate formations of the early Devonian Gribov suite, which outcrops on the anticlinal fold limbs. The fold limbs, on the south-eastern flank of the anticline, are cut by a series of faults (of oblique-slip and up-throw/thrust nature) which forms the "keyboard" structure of the cross-section of the area in question.

The deposit comprises five extensive (>600 m), banded, sheet-like layers of ore from 3-5 to 50 m thick, which are confined by lithology. The ore bodies have a monoclinal form, with small-scale folds dipping at angles of 30º-45º. The distribution of the main commodities (zinc, lead, and silver) within the ore bodies is relatively uniform. The ore body boundaries have been defined by various geophysical methods and by drilling. In plan and in axial section, the ore bodies plunge from the outcrop to the south-east, to depths of 200-350 m. Within the sheet-like ore bodies there are later, multiple ore veins and nested-veinlet segregations of ore formed by influx of later hydrothermal fluids during tectonic activity. Syn- and post-mineralization tectonic breccias do not disturb the continuity of the ore. The mineralization processes are accompanied by penetrative metasomatism in the form of silicification and dolomitization.

The main ore minerals are pyrite, sphalerite, and galena. Dolomite, calcite, and quartz are widespread as vein minerals. Chlorite, albite, micas, ruby silver, boulangerite, jordanite, arsenopyrite, cinnabar, and millerite are found in small amounts. The ores are massive (solid, nested-veinlet, breccia), vein-veinlet, veinlet-impregnated and mostly pyritic in their composition (40-90 %), the rest being sphalerite and galena of various generations. The contents of lead vary from 1.0-2.9 %, and of zinc from 1.6-20.8 %. The ores belong to the carbonate-hosted stratiform (lead-zinc) deposit type. The reserves of the Pavlov deposit, estimated by the National Reserves Committee for Categories C1 + C2 are > 2.4 million tons; the probable Category P1 resources are 7 million t, and P2, 12 million t total for lead and zinc. The silver reserves are estimated as accessory. The deposit can be developed by open-pit mining.
Figure 24: Geological map of the Norilsk District (from Strunun, 1991, see Krivolutskaya, 2016).

Deposits in the Norilsk District

Intense prospecting in the Norilsk district, related to the search for coal for the Northern Sea Route, began in 1919–1920. Norilsk was first ranked as a unique ore district in the early 1960s when the Talnakh and Oktyabrskoye deposits were discovered as a result of exploration by the Norilsk Integrated Survey Expedition. The Norilsk ore district is, in a tectonic context, an independent block with a reduced thickness of continental crust consisting of a crystalline basement and an igneous-sedimentary platform mantle. The structure of the district shows a flat to wavy distribution of an effusive-sedimentary rock series, from Cambrian to Anthropocene over 6000 m thick, and with swell-like structures, brachy-synlines and deep-seated faults, through which melts penetrated the upper levels of the section, resulting in the generation of extensive magmatic units (traps). All the rocks are intruded by dolerites and gabro dolerites of various ages and compositions. The district is at the junction of the Khantai-Rybninsky ridge and the Norilsk-Kharaelakh depression. Smaller structures have been identified within the latter: the Norilsk and Kharaelakh troughs, the Kayerkan-Pyasino and Oganer brachyanticlines and the Valkov saddle. The largest commercial deposits are located along the Norilsk-Khatanga deep-seated fault. The Norilsk and Talnakh ore clusters are located along this structure. Two types of ore are known in the deposits of the Norilsk ore district: PGE-copper-nickel-sulphide ores and low-sulphide PGE ores (Figure 24).

Norilsk 1 (PGE-Ni-Cu-Co)

The first data on the deposit is related to merchant K. P. Sotnikov who, in 1865-1868, melted 7222 pounds of crude copper out of chalcopyrite-rich argillites of the Tungusian series and sulphide veins cross-cutting them at the base of the northern slope of Mount Rudnaya. In 1920, geologist N. N. Urvantsev found primary sulphide ores. In 1922, N. K. Vysotsky who worked there by order of SibGeolCom found platinum in copper-nickel ores.

The deposit is confined to a differentiated sheet-like intrusion of the same name extending north-eastwards, with a length of ca. 12 km (Figure 25). The intrusion's thickness varies from 30-350 m, the average being ca. 130 m. The copper-nickel sulphide mineralization is related to the lower ultrabasic horizons of the intrusion (picrites, olivine gabbrros, and gabbro troctolites), near the contact, and is represented by veinlet-impregnation and massive ore types. The sulphide mineralization forms a persistent ore horizon, matching the outline of the intrusion. Veinlet-impregnated ores of the exocontact zone form an intermittent "aureole" around the intrusion. In the north, the intrusion splits into two branches, the western ("Coal Creek") and eastern ("Bear Creek") Contact metasomatic alteration in the deposit is negligible; calcsilicate skarn occurs locally in fissures. Biotitization and chloritization are widespread processes, but discordant to the structures.

The copper-nickel sulphide mineralization is represented by disseminated ores and nest-like accumulations of pyrrhotite, pentlandite, and chalcopyrite, occurring mainly in the lower, olivine-rich picrite and taxite, and in contact dolerites (Figure 25 and 26). Schlieren ore bodies have limited occurrence among the taxitic dolerites in the intrusion’s basal part. The schlieren ores contain large quantities of basic plagioclase, olivine, and pyroxene, and show a gradual transition to weak impregnation. Veins of massive sulphides occur locally at the base of the intrusion and in sedimentary country rocks: veinlet-impregnated mineralization is locally developed in the country rocks. The sulphide mineralization forms, in general, a persistent ore horizon, which matches the intrusion's outline in plan.
The massive sulphides exploited in the first years on the northern ridge of Mount Rudnaya are no longer mined. They were originally rather large (up to 200×100×20 m) lenslike bodies. A 2 m aureole of rich impregnation ores crosscutting the contact of the intrusion was observed in taxite dolerites surrounding the massive ore. The massive ore veins are mainly developed in the intrusion’s northern part. They have a complex form and sharp boundaries. They contain pyrrhotite, chalcopyrite, pyrrhotite with pentlandite exsolution lamellae, cubanite-pentlandite-chalcopyrite, rich chalcopyrite or cubanite and bornite-chalcocite. Some millerite-pyrite ores are found among the massive ores.

The reserves of the deposit’s southern part are 273,000 t of nickel, 378,000 t of copper, 12,700 t of cobalt, and about 518 kg of platinoids. Norilsk 1 was, for 27 years, the only raw mineral source for the Norilsk Mill.

Talnakh (PGE-Ni-Cu)

The Talnakh deposit is located in the Kharaelakh trough, 25 km NE of Norilsk, near the town of Talnakh. The deposit was discovered in 1960 by G. D. Maslov, V. F. Kravtsov, Yu. D. Kuznetsov, V. S. Nesterov, and Ye. N. Sukhanova, researchers of the Norilsk Survey Expedition, during prospecting work. The first mine, Mayak, started operation on April 22, 1965. Currently, the deposit is operated by underground mining at the Komsomolsky mine comprising the Mayak, Komsomolskaya and Skalistaya shafts. The deposit development license is owned by MMC Norilsk Nickel.

The copper-nickel ores of the Talnakh ore cluster, comprising the Oktyabrskoye and Talnakh deposits, are related to a large gabbro-dolerite intrusion located in the southern part of the Kharaelakh trap trough (Figure 27). The trough is intersected along its long axis by the NNE-SSW-trending Norilsk-Kharaelakh ore-controlling fault, which determines the positions of the intrusions hosting these deposits.

The deposit is connected to the interstratal, differentiated Talnakh basic-ultrabasic intrusion. The ore-bearing intrusion consists of...
several separate branches diverging from the assumed feeder. The Talnakh deposit is confined to the intrusive branches of the so-called "top" ore level located in Carboniferous-Permian terrigenous-carbonate rocks (Tungusian series), comprising the north-eastern, central, and south-western branches. The stratified magmatic bodies are divided into the following units (from the top): 1) eruptive breccias, leucocratic gabros; 2) gabbro dolerites and quartz dolerites; 3) olivine-free dolerites; 4) olivine dolerites; 5) picritic dolerites, olivinites; 6) taxitic and contact dolerites. The intrusions are surrounded by aureoles of contact hornfels about 20 m thick on average, among which are found skarn, albite-microcline metasomatites and serpentinites.

The main bulk of the copper-nickel sulphide ores is localized in the area of the lower endo- and exocontacts of the nickel-bearing massifs; some bodies of impregnated and massive ores are also located in the roof of the intrusion. The main bulk of the ores is in picrite, taxite and in contact dolerites. Impregnation has, in some cases, been recorded in rocks of gabbroic composition. Veinlet-impregnated ores are localized in the exocontact zone and are confined to altered sedimentary rocks (Figure 28). The ore bodies are sheet- and lens-like, and their outlines are similar to that of the intrusion in plan. Three types of ore are found: 1) impregnation in parent rocks, 2) massive sulphides near the base of the intrusion and 3) veinlet-impregnation in exocontact rocks. The primary ore minerals are pyrrhotite, pentlandite and chalcopyrite. Cubanite, magnetite, titano-magnetite, and ilmenite are also common. Talnakhite, mooihoekite and trolite commonly form significant aggregates in massive ores. The subordinate ore minerals include pyrite, millerite, bornite, valleriite, heazlewoodite, covellite and several PGE-bearing minerals.

The most common ore textures are massive, veinlet-impregnation, breccia, and disseminated. The most common structures are porphyritic, hypidiomorphic granular, subgraphic, drop-shaped and siderornitic. Apart from copper and nickel, the ores contain cobalt, platinum metals, gold, silver, tellurium, and selenium.

The metal reserves of the Talnakh deposit, as of 01.01.2012, were:

- Proven copper reserves: Category A+B+C1, 7,858,500 t Cu: the average content in the ores is 1.11 % Cu; inferred reserves, as Category C2, 2,761,900 t Cu;
- Platinum group metal reserves (palladium and platinum): for Category ABC1, 3,275 t Pd+Pt (4.6 g/t), for Category C2, 1,224.9 t Pd+Pt.
- Nickel reserves remain confidential, but the average Ni-content in explored ores is 0.69 %.

The Talnakh deposit is Russia's second largest in terms of nickel and platinum group metal reserves (after Oktyabrskoye which is located NE of Talnakh). Its proportion in Russia's explored nickel reserves for Category A+B+C1 was 23.6 % as of 01.01.2012. For copper reserves, the Talnakh deposit is Russia's third largest, after Oktyabrskoye and Udokan. Its proportion in Russia's commercial copper reserves for Category A+B+C1+C2 was 11.4 % as of 01.01.2012. Geological surveys are currently progressing on the northern flanks of the Taimyrsky mine, the eastern flanks of the Skalisty mine, around the Mayak mine, and on the southern flank of the Talnakh deposit. Copper-nickel sulphide ore bodies have been intersected by several drillholes. The expected growth in reserves in 2014–2016 is >500,000 tons of nickel. A geological survey is being conducted on the south-eastern flank of the Bear Creek open pit (Figure 29). The expected growth in ore available for open pit mining will (according to Norilsk Nickel) be about 40,000,000 t.
Maslov (Pd-Pt-Ni-Cu)

The Maslov deposit of platinum-palladium-copper-nickel ores is located 12 km S of Norilsk in the northern part of Krasnoyarsk Krai (Figure 30). The deposit is located in the eastern branch of the Norilsk-1 intrusion, and adjoins the southern boundary of the deposit of the same name.

The Maslov deposit was discovered in 2009 during a geological survey by MMC Norilsk Nickel. The deposit was given its name after the famous Norilsk geologist, G. D. Maslov, who indicated the prospects of the area along the Norilsk-Kharaelakh fault S of the Norilsk-1 deposit in the 1960s. The deposit contains copper-nickel sulphide ores with exploitable grades of nickel, cobalt, copper, PGEs, gold, etc. The main reserves of the Maslov deposit are concentrated in a large sheet-like layer 27-45 m thick. The layer extends along the intrusion’s central axis and has a configuration near-concordant to it in plan. The ore body lies at depths of 800-1100 m. Mineral reserves data are given in Table 1.

<table>
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<th>Metal</th>
<th>Category</th>
<th>C1+C2</th>
<th>Reserve (Mt)</th>
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<tr>
<td>Platinum</td>
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<td></td>
<td>12479</td>
<td>1.78 g/t</td>
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<td>1122</td>
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<td></td>
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</tr>
<tr>
<td>Gold</td>
<td></td>
<td></td>
<td>1305</td>
<td>0.19 g/t</td>
</tr>
</tbody>
</table>

Table 1: Mineral reserves of the Maslov deposit and metal content in ore (www.nornik.ru/).

Norilsk 2 (Pd-Pt-Ni-Cu)

The Norilsk 2 low-sulphide platinum group metal deposit (Figure 31) is located 7 km from Norilsk, also within Krasnoyarsk Krai. The deposit was discovered during a geological survey conducted by Norilskstroy under the supervision of N. N. Urvantsev in 1926.

The low-sulphide mineralization in the Norilsk 2 intrusion is confined to upper heterogeneous gabbro series rocks, which include taxite, gabbro dolerites with chromite, leucogabbros, olivine-bearing and olivine gabbro dolerites and eruptive breccias of carbonaceous argillites. The thickness of the low-sulphide mineralization is highest in the northern part of the intrusion and average in the central part. Three types of palladium-platinum-copper-nickel sulphide ores are identified in the deposit:

- Massive ores in the intrusions and their exocontacts
- Impregnation and veinlet-impregnation ores in the intrusions
- Impregnated, veinlet-impregnated and breccia-like ores in the exo- and endocontacts of ore-bearing intrusions (“cupriferous.”)

The platinum-group metal mineralization is present as palladium arsenides but also as tellurides, and platinum arsenides. The total platinum-group metal content seldom reaches 1.67 g/t, and in unique cases, 3.3 g/t, which is at the level of average values in similar intrusion horizons in Norilsk I. The sulphide associations typical for this low-sulphide mineralization is
pentlandite-chalcopryite-pyrrohite and pyrite-chalcopryite, both of which have variable proportions of the main sulphides. Currently, Norilsk 2 is considered a non-economic deposit.

**Chernogorskye (PGE-Ni-Cu-Au)**

The Chernogorskoye deposit is 15 km SE of Norilsk, also within Krasnoyarsk Krai (Figure 32). The first ore discoveries on the site were made in 1943-1944 during a detailed survey and prospecting effort. The deposit forms a sheet-like layer extending approximately E-W and with a trough-like cross section. The ore-bearing differentiated intrusion (which is up to 200 m thick) typically has a significant thickness of acid hybrid ores in the roof and a wide (up to 40 m) aureole of contact alteration (of hornblendites, magnesian and calcic skarns). The sequence of differentiated units is very persistent. The impregnated ore horizon is confined to picrite, taxite, and contact dolerites. Their composition is similar to that of the Norilsk 1 deposit. The license for prospecting, extraction, and production of sulphides and precious metals of the Chernogorskye deposit is owned by the Chernogorskaya Mining Company.

The Chernogorskaya Mining Company was acquired by Russian Platinum in April 2011. Mining (open pit) was expected to commence in 2015. Future development of the deposit by underground mining is also probable. It is intended to build a dressing mill in the vicinity of the deposit, and a smelter at which concentrate will be processed to produce copper-nickel matte. Construction of a plant for hydrometallurgical processing of matte and production of marketable metals is being considered. The approved commercial ore reserves are 143 million tons, the average content of nickel being 0.22 %, of copper 0.29 %, and of platinum group metals and gold 3.9 g/t. Platinum group metals occur at a depth of 20 m in the western part, and at 245 m in the eastern part of the ore body.

**Copper and Nickel Deposits of the Pechenga Ore District**

The Sputnik, Verkhneye, Zhdanov, Tundra and other deposits are part of the Pechenga group of deposits, located in the Eastern ore cluster of the Pechenga ore district. Administratively, this group of deposits is located in Murmansk Oblast, on the Kola Peninsula close to the Norwegian border. The deposits occur in two near-parallel levels extending for ca. 9 km in an east-south-easterly direction within the Pilgujarvi Suite of tuffaceous sedimentary rocks.

The ore-bearing differentiated intrusions and copper-nickel deposits within them are located both at the bottom and top of the suite; they are accompanied by relatively small bodies of altered peridotite and gabbro, and by gabbro-dolerite sills. The differentiated Kierdipori massif and the related Sputnik deposit are on the western flank of the ore cluster, at the very bottom of the Pilgujarvi Suite. The ore-bearing intrusions of the Verkhneye deposit are found to the south, at the top of the Suite (Figure 33); these link to the east with the Tundra deposit, which in turn is the western extremity of the Zhdanov deposit. The Zapolarnoye deposit is located in the footwall of the important Pilgujarvi nickel-bearing intrusion along the contact with dolerites. The Onki deposit forms the extreme south-eastern flank of the Eastern cluster.
The Sputnik deposit, 2.5 km W of the Zhdanov deposit, was first explored and described as the Kierdjipor deposit. An additional survey in 1973–1974 (by A. M. Dudkin, V. G. Kukharukh, V. V. Trushik) revealed a new “blind” ore body confined to a thin interstratal intrusion in the footwall of the Kierdjipori nickel-bearing massif, and later named “Main”. Since then, the associated group of ore bodies has been given the collective name, “Sputnik deposit” (Figure 34). The deposit is confined to two sheet-like differentiated massifs occurring in parallel near the surface. The two massifs are in contact along a steep thrust zone in the middle of the site below the +20 m level. The main branch of the mineralized tectonic zone extends along the base of the lower massif. Both zones link to the mineralized tectonic zones of the Zhdanov deposit to the east, and to the occurrences in Mirona and Raisoaivi to the west.

The copper-nickel sulphide mineralization is mainly concentrated in the near-basal parts of both massifs, and partially along the marked tectonic faults. Ore bodies 1 and 2 are confined to the lower part of the northern Kierdjipori massif and to the upper part of the southern Kierdjipori massif, respectively. Ore body 3 is located on the western flank, ore body 4 on the eastern flank,
and ore body 5 occurs in deep horizons. The ore bodies have a sheet-like, slightly curved shape, and strike ESE and dip 40-60° SSW. Their thickness varies from 1 to 45 m. Ore bodies 1, 2, 3, 4, and 5 contain about 80% of the deposit’s ore reserves. They all consist of impregnated serpentinites. Grades in the ore bodies have the following ranges: 0.82-1% nickel; 0.46-0.77% copper; 0.02-0.023% cobalt.

The main ore body is tabular in shape and has been traced along strike in a near E-W direction for 700 m, with a dip of 50-60° S; the body tapers at about 120 m from the surface. It has been traced down-dip for >1500 m. The ore body is confined to a thin intrusion of intensely mineralized ultramafic rock occurring along the interstratum tectonic zone 50-60 m away from the base of the Kierdjipori massif. Its thickness varies from 0.5 to 25 m. Unlike the other ore bodies, it consists of rich breccia, less common massive ore (nearly 60% sulphide), rich, densely impregnated and less impregnated ores in serpentinites in about equal proportions. Mineralized xenoliths have limited occurrence. Mineralization in serpentinites predominates in the central part of the ore body where the parent intrusion is dragged into blocks along the interstratal zone. The mineralized tectonic zone is ubiquitously filled with breccia ores forming the base and flanks of the ore body. Alternation of breccia ores and densely impregnated ores in serpentinites is observed in the central part of the orebody. Breccia ores occur moreover, not only on both sides of the mineralized intrusion but also within it. The main ore body is the only one in the deposit in which ores in serpentinites are of secondary importance. They are most widespread in the top part of the layer where they form lenses among breccia ores. The contents of economic components in the ore body are: 1.0-4.5% nickel; 0.5-2.0% copper; 0.020-0.080% cobalt.

Verkhneye Nikkel-Copper Deposit

The Verkhneye deposit is at the very top of the ore-bearing series, 700-800 m up-section from the Sputnik deposit. The ore bodies of the deposit were first intersected in 1970 by the Kola super-deep borehole SG-3 and were traced up section by the Pechenga survey crew with two drillholes at a distance of up to 1000 m from the SG-3 hole. Further exploration was suspended due to the great depth of the ore body. In 1980-1990, the same concealed ore bodies, then grouped under the name Verkhneye deposit, were explored simultaneously with prospecting of the deep horizons of the Sputnik deposit (Figure 35).

The Verkhneye deposit comprises three ore bodies confined to the near-basal parts of two superimposed, slightly curved interstratum intrusions of altered gabbro-peridotites 20-130 m thick and 600-200 m long traced to a depth of over 1700-1800 m. Nickel-bearing intrusions and
ore bodies occur concordant with the bedding of enclosing tuffaceous sedimentary rocks, and have a strike and dip, on average, at ca. 40° SW. Two sheet-like series of typical impregnated ores in altered peridotites, the Western and Eastern ore bodies, have been explored in the upper intrusion. The so-called Main ore body, also formed mostly by impregnated ore in talcose serpentinites, is confined to the lower, and longest intrusion. The strike lengths of the ore bodies are: Main, 200 m; Western, 1000 m; Eastern, 400-600 m. Their thickness is irregular, varying from 0.8 to 60 m (in swells). The Main ore body is not completely delineated, and continues even deeper than the level of its intersection with the Kola super-deep borehole, the absolute depth below the surface being 1315-1325 m. Typical impregnated ores in talcose serpentinites are predominant in the composition of all the ore bodies: ca. 2 % is rich impregnated, breccia-like and massive sulphide ore. The mineralogical composition of these ores is similar to that of the Sputnik deposit.

Zhdanov (Pilgujarvi) Ni-Cu Deposit
The Zhdanov deposit has a central position in the Eastern ore cluster. It was studied by Finnish geologists in 1929-1934, but was found then to be unviable due to the low metal content of the ore. In 1945-1946, an expedition from the Leningrad Geology Agency gave the deposit the first positive estimate. Between 1947 and 1959, the deposit was explored in detail to depths of 600-800 m by the Pechenga survey crew. Six interconnected ore bodies have been identified: the Central, Eastern, South-Eastern, Southern, Western, and South-Western ore bodies. In 1959, the deposit was transferred to the Pechenga Nickel mill for development, and open-pit mining started the same year.

The formation and localization of the Zhdanov ore bodies are closely related to the inner structure and tectonics of the main differentiated gabbro-peridotite massif, although the deposit area contains about ten more nickel-bearing intrusive bodies of similar composition but smaller size. The main massif is shaped as a curved interstratum intrusion with a total length of up to 6 km, dipping at 45-60° SW. Its greatest thickness (up to 600 m) is confined to a trough-shaped depression in the centre of the deposit (Figure 36 A). The massif is surrounded by intensely folded tuffaceous sedimentary rocks, and occurs in general concordantly with these rocks, reflecting the area’s folded structure in its morphology. Relatively flat-lying, exposed cross-folds, with their axes oriented south-westwards, in the direction of the common dip of the ores, control the shape and inner structure of the massif, and that of all the ore bodies. Transverse folding appears most intensely on the western limb of the deposit, where the Western, Seventh and other insufficiently studied ore lenses and small massifs are confined to the axes of synclinal folds. The massif’s inner structure is marked by layered rocks and ores consisting of (from base to roof): 1) mineralized serpentinitized olivinites and serpentinized, primary impregnated sulphide ores; 2) serpentinitized peridotites (without ore); 3) pyroxenites; 4) gabbros. The latter makes up about 65 % of the massif’s volume, whereas the proportion of serpentinitized peridotites make up about 30 %, and that of pyroxenites, 5 %.

The layer of mineralized serpentinitized olivinites and peridotites is in general concordant with the multiply-folded basal contact of the massif. The Zhdanov ore bodies are thus simply mineralized lowermost layers of the initially layered intrusion. The thickness of the primary impregnated ores with economic concentrations of nickel and copper sulphides reaches its maximum values in the central depression of the massif. The Central ore body is confined to this part of the intrusion. In anticlinal hinges on both its flanks the content of mineralization decreases to complete absence. The same ore layers continue north-westwards and south-eastwards in other synclinal depressions, but structurally slightly above the base of the central depression, forming the Western, Eastern, South-Eastern and Southern ore bodies respectively. Unlike these, the South-Western and Seventh ore bodies, and the Tundra deposit, which form the deposit’s western limb, are most probably related to tectonically separate parts of the Main massif.

The primary magmatic features of the deposit were noticeably affected by tectonic faults in the epigenetic stage. Among them, steep interstratal tectonic zones play an important role in structuring the deposit. The main tectonic zone extends along the parent intrusion’s lower contact and controls the occurrence of rich breccia and massive sulphide ores in the Central, Western and Eastern ore bodies. It extends on
the deposit’s western flank, with the same strike, to the Sputnik-Kierdjipor-Mirona and Raisoaivi deposits. Eastwards from the main zone, a straight tectonic fault extends at a low angle, controlling the occurrence of the massive sulphide ores of the Zapolarnoye deposit. Morphologically, the Main zone is defined by brecciation and foliation of mineralized serpentinitized peridotites, and in places by mineralized tectonic breccia and massive sulphide ores.

The second (Upper) interstratal tectonic zone extends parallel to the Main zone, 200-500 m up section, and dips at 70-80° SW. It splits the Pilgujarvi parent massif into two large blocks: the Main block, with the Western, Central and Eastern ore bodies located at its base, and the Upper block, which controls the positions of the South-Eastern and Southern ore bodies. A third tectonic thrust zone extends parallel to the first two, 300-400 m to the S or higher in the section, above the South-Eastern and Southern ore bodies, and controls the localization of the Third Bystrinskoye ore bodies (Figure 36 B). A fourth interstratal tectonic zone intersects the very top of the Pilgujarvi suite, 300-400 m from overlying dolerites in the nappes. Only thin lens-like, stratified bodies of ultrabasic rocks are found within this zone, with mineralization occurring locally. The vertical displacement of the blocks differs for the various areas. The maximum displacement (300-500 m) is observed in the centre of the ore cluster, decreasing somewhat on both flanks.

The general outline and relative positions of the ore bodies of the Zhdanov and other deposits of the Eastern ore cluster may be seen in Figure 36 A. Their shapes are mostly sheet-like, tabular, and lens-like, and they are equally extensive.
in strike and dip. The ore bodies have a strike length of 200 to 1800 m (Central ore body), the dip is on average 50-60°S to SE and their thicknesses vary from a few metres to 100 m.

**Fedorov Tundra Deposit (PGE)**

The Fedorov Tundra deposit is 80 km E of the city of Apatity, and 58 km SE of the district centre Lovozero. It is localized within a stratified intrusion of the same name, in taxitic norites and gabbro norites of the marginal zone, and in overlying stratified norite and gabbro norite zones, in which sulphide mineralization occurs both disseminated and as sulphide-enriched lenses. The Fedorov-Pana stratified intrusive massif forms an elongate range of hills in the central part of the Kola Peninsula. From structural and tectonic aspects, this district is a part of the northern contact between igneous-sedimentary rocks of the early Proterozoic Imandra-Varzuga paleorift structure and formations of the Archean basement.

The Fedorov-Pana massif has a sheet-/lopolith-like shape, and is exposed at the present erosional level by fragments of the northern part of the lopolith. It extends in a NW direction for nearly 80 km, with outcrop widths of 600 m to 5-7 km and thicknesses of ca. 3.8-5.0 km (Figure 37). It dips 30-50°SW in the west and up to 50-80°SW in the centre; in the south-east, the dip decreases with depth. The Fedorov-Pana massif can, according to gravimetric and magnetic data, be traced for a considerable distance under the sedimentary-igneous rocks of the Imandra-Varzuga rift-induced structure, to a depth of 4-5 km, retaining its south-westerly dip.

The Fedorov-Pana intrusion is a peridotite-pyroxenite-gabbro-norite complex. The following age data have been determined for the intrusive rocks: gabbronorites, 2501±1.7 Ma and 2491±1.5 Ma; gabbro pegmatites, 2470±9 Ma; anorthosites, 2447±12 Ma. The massif has a limited differentiation sequence as it has no lower ultramafic and "critical" zones. It consists of a poorly differentiated series of mafic rocks represented by gabbronorites. This type of structure is, however, quite typical for layered intrusions in the Baltic Shield.

The intrusion is divided into large blocks by transverse fault zones. From west to east, the following have been identified: The Fedorov block, mostly consisting of gabbroid rocks, and separated from the West Pana block by the thick Tsaga fault zone; the Lastyavr block, adjacent to the fault zone, and intensely deformed; the West Pana block, the thickest, most well-exposed and well-studied part of the intrusion, consisting mainly of gabbronorites, within which two

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*Figure 37: Geology of the Fedorov-Pana intrusion (Karpov, 2004, Pana JSC). 1) Gabbro zone (GZ); 2) Gabbro norite zones (GNZ 1, 2, 3); 3) Upper stratified horizon (USH); 4) Lower stratified horizon (LSH); 5) Stratified horizon of Eastern Pana block Norite zone (N2); 6) Lower marginal zone (LMZ); 7) Bodies of magnetite gabbroros; 8) Gabbronorites. 9) Belaya Tundra complex of alkali granites; 10) Plagioclase gneisses and schists; 11) Amphibolite; 12) Igneous and sedimentary formations of the Imandra-Varzuga series; 13) Tsaga massif of gabbro labradorites; 14) Quaternary sediments, moraine.*
stratified horizons stand out: the “upper” (USH) and “lower” (LSH) Stratified Horizons. The East Pana block has a heterogeneous lateral structure, with a predominance of gabbros in the section, and a stratified horizon in the section’s basal part which resembles the structure of the stratified horizons in the West Pana block. The main PGE-enriched levels in the Fedorov-Pana massif are related to the stratified horizons. Known PGE ores in the intrusion are of two types: A near-basal layer of low-sulphide copper/nickel/PGE ores in the Fedorov tundras massif; and low-sulphide PGE reefs in the western and eastern part of the Pana massif.

The Fedorov tundra massif shows the sequence: Marginal norite (taxite gabbronorite) and plagiopyroxenites, gabbronorite and gabbro zones. The marginal-zone, so-called taxite gabbronorites extend to near the base of the body. The norite-plagiopyroxenite zone, occurring higher in the section, reaches a thickness of 200 m. The largest part of the Fedorov block section is formed by equally thick units of gabbronorite and massive gabbro. The gabbros contain rare, thin streaks of leucocratic gabbro and anorthosite. The platinum metal mineralization is inclined to the marginal and norite zone for a total width of up to 1 km in plan, traceable for 10 km. A feature of this massif’s platinum metal occurrences is levels with almost equal contents of platinum and palladium (up to 3.53 and 3.91 g/t respectively), but with platinum predominating in some parts. The main platinum-group element minerals found are: kotulskite, braggite, and merenskyite. Less common precious metal-bearing minerals include moncheite, vysotskyite, sobolevskite, stillwaterite, sperrylite, and gold.

The approved reserves are: A+B+C1: 173 Mt of ore (238 t of PGE); C2: 87 Mt of ore (109.9 t of PGE); non-economic reserves are approved at 15,823,000 t of ore (18.5 t of PGE). It is intended to develop the deposit in two open pits with a total annual capacity of at least 12 million tons of ore. The ore will be processed by flotation. The mill’s finished product will be a concentrate containing copper, nickel, gold, platinum and palladium for further metallurgical processing and refinement.

The Kievei PGE deposit

The Kievei PGE deposit is situated 120 km E of Apatity, and was discovered in the period 1991-1999. It is related to a mafic-ultramafic complex in the central part of the West Pana massif, a layered intrusion of the Fedorov-Pana tundras (see Figure 37). The massif is divided into the upper and lower stratified horizons: the Kievei deposit is confined to the lower horizon (LSH for short). The PGE mineralized level is called the North Platinum Reef. The LSH is found 600–800 m above the base of the massif. It is marked by thin layers of norite, pyroxenite, spotty leucocratic gabbro and anorthosite among alternating gabbronorites and gabbros with consortal (extensively intergrowing) texture. The average thickness of the LSH is 46 m, with a dip of 30°S.

Sulphide and platinum-group metal mineralization is observed in the bottom and middle parts of the LSH section. The ore body thickness is 3–15 m. The ore mineralization occupies interstices between grains in all the rocks of the lower stratified horizon. The thickness of the sulphide-bearing rock varies from a few centimetres to 2–3 m. Two layers of ore have been identified; the Main and, less persistently, the Upper layer. The layers consist of impregnation which has no distinct boundaries: the outlines of the ore are therefore based on sampling results. The ore is heterogeneous: PGE-bearing mineralization is confined to the boundaries of the spotty leucocratic gabbros and anorthosites with norites and gabbronorites, with plagiopyroxenites and rarely with plagi-websterites. The main ore-bearing zone comprises several contiguous sheet- and lenslike sulphide-bearing layers occurring en echelon and with lengths of up to 100-400 metres. The distribution of sulphides in the mineralized layers is irregular: the sulphide content varies from <1% to 3.5 %. The thickness of the main ore body varies from 0.1-6.5 m, averaging 1.7 m. The approved reserves (1.01.2009) were: A+B+C1: 1.8 Mt of ore giving 6.5 t of PGE.

The main economic elements are palladium, nickel, platinum and copper. The average content of platinum is 0.479 g/t, and of palladium 3.172 g/t (Pd/Pt-ratio 6.7). In low-sulphide ores, the primary minerals are chalcopyrite, pentlandite and pyrrhotite. Apart from pentlandite, palladium is present in kotulskite, vysotskyite,
merenskyite and braggite, and platinum in moncheite, braggite, merenskyite and vysotskite. Some gold is present as electrum. The platinum minerals are generally associated with the sulphides. The morphology of the platinum mineral grains is diverse. Both automorphic crystals of PGE minerals and xenomorphic grains and limbate (zoned) segregations are observed. The internal structure of grains is blocky and zoned. PGE arsenides occur locally, leading, because of alteration, to an increasing proportion of sperrylite.

**Vuruchuaivench (PGE)**

The Vuruchuaivench platinum-group metal deposit is located in the central part of the Kola Peninsula, 10 km from Monchegorsk and 5 km from the operating Severonickel mill of Kola MMC (Norilsk Nickel). The Vuruchuaivench site was
discovered by KOLAMMC surveys in Monchegorsk District in 2004-2008, and is now being prospected for precious metals. The Vuruchuaivench mineralization (Figure 38) is a reef layer with a horizon of disseminated sulphide enrichment in albitized anorthosites in the western and central blocks of the ore body, and in leucocratic metababbronorites in the eastern block. According to prospecting data from 2004, the Vuruchuaivench reef extends in a SW-NE direction for 1.0–1.2 km as a layer with individual attenuating lenses. The Vuruchuaivench massif adjoins the norite-gabbronorite part of the Monche-Pluton section (Mount Nyud-Poaz) in the south. It is a wedge-like body, dipping gently south-eastwards, resting on Archaean granite gneisses and overlain by Imandra-Varzuga series rocks, under which the massif is traceable southward in drillholes. The massif’s rocks crop out SW and SE of the Nyud-Poaz massif, and are traceable for 7-8 km in a north-easterly direction, before disappearing under the Lake Imandra basin. The deposit is broken up into blocks by a series of thrusts and faults. The ore bodies are represented by sheet-like beds and flattened lenses without clear outlines. The main reserves are concentrated in Ore Body No. 1, and confined to a layer 0.5 to 10–20 m thick with sulphide impregnation.

The Vuruchuaivench massif consists of two main rock types:

- Melanonorites, norites and gabbronorites, the upper part of which underwent post-crystallization alteration, form the lower part of the section

![Figure 39: Cross section of the Peschanka porphyry deposit (after Migachev et al, 1995, in Seltmann et al, 2010).](image-url)
• Talc-chlorite (zoisite-vuagnatite)-plagioclase rocks, previously described as meta-leucogabbros, meta-anorthosites or plagioclases – form the upper part of the section.

Data on the commercial reserves of the Vuruchuaivench deposit are presented in Table 2.

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<thead>
<tr>
<th>Reserves, Vuruchuaivench deposit, Category C1+C2</th>
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<tr>
<td>Ore (Mt)</td>
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<td>Nickel (kt)</td>
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<td>Gold (t)</td>
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Table 2: Reserves of the Vuruchuaivench deposit. (Norilsk Nickel Annual Report 2014. p 65.)

Peschanka (Cu-Au-Mo)¹

The Peschanka (Baimskoe) Cu–Au–Mo porphyry deposit (Figure 39) (Migachev et al. 1995) contains resources of 1350 Mt of ore grading 0.61% Cu, 0.015% Mo, 0.32 g/t Au and 3.7 g/t Ag corresponding to 8.3 Mt Cu, 200 kt Mo, 425 t Au and 5 kt Ag. The deposit is situated in the Circum-Pacific orogen (Chukotka Peninsula), close to the western margin of the Okhotsk-Chukotka volcanic–plutonic belt, where this belt is superimposed on the Omolon microcontinent. The belt contains multiphase diorite–monzonite–syenite–granite intrusions, located within the Byam metallogenic zone. This metallogenic zone forms a linear structure traceable for over 200 km, with a width of 20–25 km, and hosts a number of porphyry deposits. Monzodiorite and quartz monzodiorite intruded by younger granodiorite porphyry stocks are present in the deposit area. Hydrothermal alteration has created an innermost zone of quartz–sericite–carbonate assemblage, an intermediate zone composed of K-feldspar and biotite, and an outer zone composed of propylitic (hydrosericitic) assemblages (Figure 39). The Cu porphyry mineralisation is coincident with the zone of quartz–sericite–carbonate alteration replacing the entire granodiorite porphyry body, whereas the Mo mineralisation is confined to a narrow area inside the Cu halo, generally associated with a dense quartz stockwork. The mineralization formed during four stages (from earlier to later): (I) molybdenite; (II) chalcopyrite–pyrite; (III) bornite–chalcopyrite–tennantite; and (IV) chalcopyrite–sphalerite–galena stages. The chalcopyrite–pyrite mineralisation is most widespread, but the higher-grade ores are associated with the bornite–chalcopyrite–tennantite assemblage.

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¹ Description taken from Seltmann (2010)
Aganozero chromium deposit

The Aganozero chromium deposit is located in Pudozh District, Republic of Karelia, 30 km E of Lake Onega and 40 km N of the city of Pudozh, near Lake Aganozero. The first data on the chromium ores in this area were obtained by the Karelian integrated Geological Survey Expedition in 1956.

Structurally, the deposit is confined to the northern (Aganozero) block of the Burakovos-Aganozero layered mafic/ultramafic intrusion (Fig. 40). The intrusion has been traced for 7100 m along strike: its thickness is 110-150 m. Titanium-magnetite mineralization is present as sheet-like ores at the base of the intrusion and is traceable to depths of 330.7 m. The vertical thickness of the layer varies between 7.4 and 32.2 m. Individual ore lenses occur higher in the section. The chromite mineralization is confined to the top of the peridotite subzone in thin streaks and lenses, and to the contact between the ultramafic and pyroxenite zones, the main chromite horizon in the massif’s layered series. In all, eight chromite horizons, 0.1 - 9.0 m thick, are known in the section, but of these, only the main chromite horizon is of commercial importance. The ore occurs in pyroxenites and poikilitic peridotites of varying composition. Olivine clinopyroxenites or wehrlites occur at the top of the main chromite horizon, and poikilitic wehrlites, or, less commonly, lherzolites and harzburgites, occur at its base. The chromium ores typically have an olivine-pyroxene-chromite-spinel composition with a fine-grained texture and two generations of chrome-spinel. The main chromite horizon forms a synform extending 8.5 km in an approximately N-S direction and up to 3.6 km wide, with the stratum dipping towards the structure’s centre at 11-33° in the southern, northern, and western parts of the deposit, and at 30-53°, in the eastern and south-eastern parts. The synform is divided by thrust faults into several tectonic blocks with vertical displacements of 20 - 170 m relative to each other. The main chromite horizon is a layer, the internal structure of which is locally complicated by thin (0.2-1.2 m) streaks and lenses of unmineralized rock – wehrlite and, less commonly, clinopyroxenites, with lengths from a few metres to hundreds of metres. On the whole, the main chromite horizon shows a persistent dip and strike. Its true thickness varies from 0.7 to 6.3 m, with average thicknesses of 3.2, 2.6 and 2.3 m for cut-off grades of 10, 15 and 20 % Cr$_2$O$_3$ respectively. The thickness of the ore increases along strike and at depth in the N-S direction. The host rocks are pyroxenites and poikilitic peridotites with differing structures.

The reserves and inferred resources of the Aganozero deposit are, in total, 205 Mt with an average of 21.78 % Cr$_2$O$_3$, which is considered...
to be a large deposit. The size of the deposit is comparable to that of the Kemi deposit in Finland, and exceeds those of the Sopcheozero and Bolshaya Varaka deposits in Murmansk Oblast. Open pit mining is possible to depths of 40-60 m, according to a feasibility study, on a relatively small part of the deposit (in the marginal parts of the synform). Underground mining is economic in the remaining part. Sparsely impregnated, medium-impregnated and densely impregnated ores are distinguished on the basis of their chrome spinel content, which reaches 75-85 %. The chromium ores contain accessory platinum metal and gold mineralization. The weighted average concentrations are: for platinum, 0.087 g/t, for palladium, 0.12 g/t, for rhodium, 0.001 g/t, and for gold, 0.25 g/t. The commercial chromium reserves to a depth of 220 m are, for Category C1+C2, 26.5 Mt, of which 7.8 Mt for open pit mining (to a depth of 50 m) and 18.7 Mt for underground mining.

North Urals Bauxite Deposits (Kalyinsky, Novo-Kalyinsky)

The North Urals bauxite district is located in the western part of the Nizhny Tagil synclinorium. The bauxite ores are confined to the monoclinal zone of the western end of the Shegultsan syncline comprising Silurian and Devonian limestones. The whole complex of rocks strikes N-S and dips 18 – 40° E. The bauxite deposits are of the karst sheet-like type. The bauxite basin is divided by faults, trending approximately E-W, and by sterile intervals into several large areas that were conventionally called deposits: Krasnaya Shapochka 7.7 km long (with the Zavagransky, Main Ore Field, First, Second, and Third Northern, and Eastern Shoot sections); Kalyinsky deposit 5.5 km long (South Kalyinsky, Central Kalyinsky, and North Kalyinsky sections); Novo-Kalyinsky deposit 2 km long; Cheremukhovsky deposit 7.9 km long (with the South Cheremukhovsky and Kedrovsky sections); Sosvinskoye deposit, and Vsevolodoblagodatskoye deposit.

The geology of the North Urals deposit area comprises Upper Silurian and Devonian formations (Fig.41). The Wenlock (Silurian) is represented by andesite-basalt porphyries (Pokrovskoye Suite). Higher in the section there are formations of Ludlow age, which in their lower part are made up of massive pink limestones (Voskresenskoye Suite) and bedded, dark-grey limestones (Kolonga Suite), and in the upper part, by dolerite porphyries, conglomerates, tuff breccias, and tuff shales with streaks of grey limestone (Sosva Suite).
The Subrovsky horizon of bauxites is directly underlain by poorly defined sediments of Ludlow and Pridolian age. In the lower part, there are conglomerates, sandstones, shales, and dark-grey bituminous limestones (pink and light-grey) of the Petropavlovskaya Suite.

Bauxite ores of the SUBR horizon (held by North Ural Bauxite Mine) occur on the uneven karst surface of the reef limestones of the Petropavlovskaya Suite. The bauxites are overlain by limestones of Eifelian age, which are divided into three units (from the base upwards): dark grey, bituminous, amphibole-bearing, bedded; light grey, massive, reef; dark grey flag-like and thin flag-like, with streaks of clayey and clayey siliceous shales. Limestones of Givetian and Frasnian ages occur higher in the section.

The ore horizons have a complex, sheet-like shape determined by the relief of the underlying limestones, post-mineralization tectonics and karstification processes. Bauxites develop in relation to erosion, and occur locally, unconformably, on pink karstic limestones of the Petropavlovskaya Suite, where they are abundant in large, irregularly shaped pits and sinks, 1 - 25 m deep, so that the ore thicknesses vary correspondingly. The strata vary in thickness from 0.1 - 9 m, with an average of 2 - 5 m. The strike length commonly exceeds 3 – 3.5 km, and the ores have been traced to depths of more than 1600 – 2000 m.

The bauxite horizon consists lithologically of three subhorizons, from the base upwards: red stained ores, red unstained, jasper-like ores and green-grey variegated ores. The bauxites fill the irregularities of the karst paleorelief surface. The middle sub-horizon ores are commonly found on an uneven surface of red-stained bauxites, and thinner ore bodies may consist only by them. Coarse pisolitic bauxites and bauxite conglomerates belonging to that subhorizon are widespread in all the deposits in the basin. The total thickness of red ores of the basal and middle subhorizons varies from 2 to 7 m; swells 30-40 metres thick are found, but are not common. Red jasper-like bauxites also occur as streaks and lenses in the upper subhorizon of grey bauxites. The upper subhorizon of greenish-grey variegated ores is represented by bedded bauxites, with a granular, less frequently dense, jasper-like structure. Bauxites in this horizon have a spotty ochre and brown colour. Below this level, their colour is greyish-green. Variegated bauxites occur, overlying red bauxites, with rather even and commonly abrupt contacts. The subhorizon’s thickness is 0.2 – 0.4 m, less frequently 2-3 m. Breccias of a sedimentary-metasomatic nature consisting of underlying limestone fragments cemented by bauxite are commonly found at the base of an ore body.

Red-stained and non-stained bauxites form about 80 % of the whole ore horizon. These bauxites are of the highest grade. Diaspore is the main mineral in the ore, whereas boehmite and chlorite are subordinate. Minerals containing impurities, including silica, carbon dioxide and sulphur, occur in small quantities and are mainly confined to shear fractures as pyrite and calcite. Red jasper-like bauxites are slightly inferior in quality to the other varieties due to their high silica content. Boehmite is the main mineral in this ore type; diaspore and kaolinite are subordinate. Variegated bauxites are very heterogeneous. Diaspore and boehmite are the major phases, whereas pyrite, siderite, haematite, chlorite, and kaolinite are subordinate.

The high-quality red bauxites have the following chemistry: \( \text{Al}_2\text{O}_3 \) 53-55 wt.%, \( \text{SiO}_2 \) 6 wt.%, \( \text{Fe}_2\text{O}_3 \) 23-25 wt.%, \( \text{CaO} \) 1.6-2.5 wt.%, \( \text{SO}_2 \) 0.4 wt.% (up to 1.1 %), \( \text{CO}_2 \) 1.9 – 3.6 wt.%, \( \text{TiO}_2 \) 2.0-2.5 wt.%. High-sulphur pyrite and variegated bauxites containing 1 - 15 wt.% of sulphur make up 5 % of the overall reserves. The North Urals bauxites are enriched in REE and contain 19 - 57 g/t scandium, 4.8 g/t yttrium, and 1.55 g/t ytterbium. The deposits are being mined by Sevuralboksitruda JSC: shafts in the underground mines reach 1000 – 1500 m.

Keiv Deposits of High-Alumina Raw Materials

High-alumina kyanite deposits are located in the eastern part of the Kola Peninsula, within the 200 km long, northwesterly striking Keiv ridge. The geology of the ridge comprises deeply metamorphosed sedimentary rocks of the Lower Proterozoic Keiv series, which have been partly metamorphosed to kyanite-rich schists. The kyanite deposits were discovered in 1928 during expeditions of the USSR Academy of Sciences.
The eastern deposits of the Keiv suite were discovered by A. A. Grigoriev, and those in the west, by A. A. Vorobyova. Exploration activities in this district have revealed 29 kyanite deposits (Fig. 42), of which the largest are Shuururta, Bezmyannoie, Tyapysh-Manyuk, Vorgelurta, Novaya Shuururta, Chervurta and Bolshoi Rov. The proven reserves of these deposits exceed 3 billion tons. The use of kyanite and other aluminosilicate minerals as a potential source of alumina was developed in the Soviet Union, due to the low grade of known domestic bauxite resources (Shabad, 1976; Voytekhovsky & Neradovsky, 2012). The Keiv series of rocks forms a large synclinorium with a general westerly / north-westerly strike complicated by multiple smaller folds, structures which are found even on a micro-scale. The synclinorium’s northern limb is overturned southwards, so that a reverse stratigraphic sequence is observed. The rocks dip at 60-80° N. The synclinorium’s southern limb shows a normal stratigraphy; the dip is also northwards but at lower angles (20 - 40° N). The Keiv series rocks are subdivided into two main parts, a lower gneiss part and an upper schist part. The latter is divided in six identifiable units (Nosikov, 1960).

All known kyanite deposits are confined to unit 6 in which three main kyanite schist types are identified: 1) Radiate/fibrous, 2) paramorphic and 3) concretion ore (Fig. 43).

Paramorphic kyanite schists (Manyuk type) are characterised by white kyanite, mostly as pseudomorphs of andalusite: a cross typical for chiastolite can be seen in their cross-section. The most interesting deposits of the Manyuk type are Vorgelurta, Manyuk, and Bezmyannoie.

**Vorgelurta, Manyuk and Bezmyannoie kyanite deposits**

The Vorgelurta deposit is located on the north-eastern slope of a tundra of the same name, 20 km NW of the Sukhoi River and 180 km E of Apatity Station on the Kirov Railroad. Gneisses, garnet-mica-, kyanite-, paramorphic kyanite- and staurolite-kyanite schists belonging to the Proterozoic Keiv series are the main rock types in the area of the deposit. Structurally, the deposit is confined to the northern limb of a synclinorium which is overturned southwards, and which is complicated by secondary and tertiary folds.

Morphologically, the deposit is represented by two sheet-like bodies of paramorphic kyanite schists, 35 and 50 m thick, separated by mica-quartz and mica-staurolite-garnet schists.
Kyanite pseudomorphs in the schists reach a length of 20 cm along the major axis, with cross-sections of up to 5-6 cm. The bodies have been traced for 5 km along strike. The kyanite content in individual samples varies from 32.75 to 45.5%, with an average of 37.93% for the deposit as a whole. Two other deposits of paramorphic kyanite ores, Manyuk, and Bezymyannoye, are located respectively 275 km and 200 km E of Apatity Station on the Kirov Railroad: these differ from the Vorgelurta deposit by the smaller grain size of the kyanite pseudomorphs (for chiastolite), with lengths up to 5-7 cm, and cross sections of 2-3 cm.

Tyapysh-Manyuk, Tavurta and Novaya Shuururta kyanite deposits

Concretion kyanite schist deposits are characterised by spherical and, less frequently, spindle-shaped aggregates, consisting of black, fibrous or acicular kyanite crystals, tightly clustered and resembling a concretion. The size of the concretions varies from several millimetres to 5-7 cm across; spindle-shaped concretions reach 10-12 cm along the major axis. Concretion schists commonly grade into pseudomorphic varieties along strike (Bezymyannoye and Vorgelurta deposits). Transitions from concretionary kyanite schists into divergent and fibrous types were observed in two cases (the Novaya Shuururta and Tyapysh-Manyuk-Tavurta kyanite deposits). Currently, three deposits of concretion-type ores are known: Tyapysh-Manyuk, Tavurta, and Novaya Shuururta.

The Tyapysh-Manyuk and Tavurta deposits are located 12 km W of the Bezymyannaya tundra deposit and 200 km E of the Apatity Station on the Kirov Railroad. The deposits were explored by the North-Western Geological Agency in 1950-1953. The deposits are confined to the northern limb of the Keiv synclinorium which is overturned southwards. The rocks dip northwards at up to 80°. These kyanite schists contain large- and small-concretion varieties with average kyanite contents of 30-32%. Small-concretion kyanite schists are difficult to process, therefore only large-concretion varieties can be regarded as ores in the deposits. The latter occur as sheet-like bodies traceable within the Tyapysh-Manyuk deposit for almost 4 km, and within the Tavurta deposit for 1.4 km.
The thickness of the ore bodies is variable, partly due to replacement of large concretion kyanite ores by other types of kyanite schist. Small-concretion schists are primarily developed on the footwall side of large-concretion schists. Less frequently, they occur as lenses inside a main ore body formed by large-concretion schists.

The Novaya Shuururta deposit (Fig. 44) is located on the Shuururta tundra summit, 3 km NW of the top of the Tsitsnykury-Nollourty tundra, 2.5 km N of Lake Semuzhye, and 225 km E of Apatity Station on the Kirov Railroad. The deposit lies 95-100 m above Lake Semuzhye. The ore is large-concretion kyanite schist (as found in the Tyapysh-Manyuk deposit) and was discovered in 1952 and explored in single trenches. The average thickness of the deposit is 148 m and its kyanite content is 42.44%. East of the deposit, large-concretion kyanite ores grade via flattened (squeezed) varieties into fibrous, variably sheaf-textured kyanite schists. To the west, large concretions are replaced by smaller ones and are gradually displaced by this texture.

Deposits of divergent and fibrous kyanite schists feature kyanite as elongate needles or fibres, commonly forming diverse fibrous aggregates resembling a sheaf, a starlet or other shapes. Divergent kyanite schists are widely developed in the southern wing of the Keiv synclinorium and in the southern fold structures, both secondary and tertiary. Features of this schist type are their wavy structure and small folding. In the northern wing of the main synclinorium, divergent and fibrous kyanite schists are absent or subordinate to other kyanite ore types.

Many deposits of divergent kyanite schists have been discovered in the Keiv ridge, notably Chervurta, Bolshoi Rov I, Bolshoi Rov II, Kyrpurta, Lysturta as well as other deposits.

mostly confined to the lower productive horizon of Mass and located mainly on the southern wing of the main Keiv synclinorium or on the smaller southern folds.

Chervurta kyanite deposit

The Chervurta deposit, 26 km SE of the former Semiostrovsky Pogost village and 200 km E of Apatity Station of the Kirov Railroad, is located on the Chervurta tundra which rises 60 – 80 m above the surrounding valley. The deposit is confined to the main synclinorium’s southern wing, within which the rocks dip at 40-45° NE. The ores are kyanite schists in the basal horizon of the productive level, with a kyanite content of 38 - 40 %. The kyanite grains are grey and black in colour, and have a fibrous habit; the grains commonly form radiating and sheaf-like aggregates. The fibre length is 3.5—5.0 mm, locally more. The deposit has been explored in an area located near the highest part of the Chervurta tundra and covering part of the stratum for a length of 1.5 km. The thickness of the stratum is 38-45 m.

Iksa (North Onega bauxite district)

The Iksa bauxite deposit was discovered in 1949, and explored in 1950-1963, in particular the western parts of the Belovodskaya and Yevsyukovskaya beds which have the most favourable mining conditions. In 1965, the deposit was allocated to the North Onega bauxite mine and mining began in 1974. The deposit is located in Plesetsk District, Arkhangelsk Oblast near Navolok village, 5-6 km SW of Navolok Station. The ores are located on both banks of the R. Onega in the estuarine part of its tributary, the R. Iksa.

The deposit is confined to the Iksa basin, a secondary structure of the ancient, buried North Onega trough. The basin is 18 km long, trending in a north-westerly direction and about 7.5 km wide in the centre; its deeper part and margins have depressions carrying the main bauxite layers of the deposit. The bauxites occur, stratigraphically, at the base of the Visean section. The deposit is made up by six bauxite beds: Belovodskaya, Yevsyukovskaya, Chirkovskaya, Kudryavtsevskaia, Tarasovskaya and Kazakovskaya. The total area of the bauxite is 120 km², of which ore-quality bauxite covers 35 km², with a thickness of 0.8 - 16 m (the average thickness of the beds is 2.5-8 m). The depths of the beds are, on average, 45-75 m. The bauxite quality within the beds and in the deposit in general, is marked by a high degree of continuity.

The western parts of the Belovodskaya and Yevsyukovskaya ore beds are classified as commercial, while the eastern part of the deposit which is over lain by limestones is non-commercial. The Belovodskaya bed is the largest (over 81 % of proven reserves, of which 95 % is commercial): Yevsyukovskaya is the second largest (11 % of proven reserves, of which 70 % is commercial). The total reserves of the other four beds are about 8 %, and they are all considered non-commercial. The Belovodskaya bed is, on the basis of mining and hydrogeological conditions, subdivided into the Western, Eastern, and Zaluzhemsky fields. The Western field is somewhat elongated in an E-W direction (3.5 × 2-2.5 km) and has an area of 12,912,000 m², a thickness 1.0 - 16.0 m (average: 8.0 m) and depth 40-55 m (average: 48 m). The overburden consists of loose sandy/clayey fluvioglacial sediments.

The Belovodskaya ore body has a convex shape and thickness of 5 m or more in the central part, decreasing to a minimum economic thickness on the sides. The highest-quality bauxites form the core of the ore bed, being replaced towards the base and roof and towards the periphery by low-quality ores and by clays. The western part of the Yevsyukovskaya bed covers an area of 5,565,000 m², the average thickness of the bed is 2.8 m and it is found at a depth of 45 m. The bauxite deposit consists of kaolinite-boehmite and locally, kaolinite-gibbsite-boehmite. The deposit contains, on average, 30-40 % kaolinite, 25-35 % boehmite and 10-15 % gibbsite. In addition the deposit contains goethite or haematite as iron-bearing phases and leucoxene, rutile and anatase as titanium phases. Small quantities of calcium and magnesium oxides are related to epigenetic siderite and zeolite formation, and sulphur is bound in gypsum. On average the Cr₂O₃ content is 0.60 % but reaches 7.3 % in some areas. According to VAMI (the Russian National Aluminum-Magnesium Institute), chromium occurs as a free hydroxide adsorbed to clayey minerals, and in chrome spinel.
Compared to other deposits, the Iksa ores have high concentrations of gallium (62 g/t), V₂O₅ (1200 g/t), Cr₂O₃ (0.58-0.60 %), Sc (90-100 g/t), Li (160-450 g/t), yttrium (200 300 g/t) and REE (up to 1000 g/t): precious metals are also found.

Vezhau-Vorykva bauxite deposit (Middle Timan bauxite district)

The first bauxite stratum was found in the Vychegda River in the Middle Timan in 1949. This initiated detailed prospecting and in 1970, high quality bauxites were found in the Vorykva River, leading to discovery of the economically viable Vezhau-Vorykva bauxite deposit (Fig. 45). Later prospecting revealed new deposits and occurrences (Upper Schugor, Eastern, Zastrovskoye, Volodinskoye and Sventlinskoye), which made Middle Timan the leading bauxite district in Russia.

The Vezhau-Vorykva bauxite deposit is the largest in the whole Timan district. It contains 56.4 % of the bauxite reserves of the Vorykva ore cluster, and nearly 12 % of Russia’s commercial reserves. The deposit includes three beds: Central, Western, and Upper Vorykva beds whose parameters are presented in Table 3.

The ore bodies have a bed to lens-like shape with a relatively even, slightly convex roof and a wavy base complicated by pockets and pit-like cavities. The average chemistry on a dry basis is (%): SiO₂ 8.01; TiO₂ 2.73; Al₂O₃ 48.69; Fe₂O₃ 27.87; CaO 0.36; S_total = 0.02. Silicon module (Al₂O₃ / SiO₂): 6.08 (The silicon module is a measure of the quality of the bauxite, the higher the value the better the quality, Nikolaeva et al., 2015). In the Central bed, three blocks of stained, low-iron bauxites which are suitable for production of abrasives, high-quality refractories and ceramics for electrical applications have been exposed. The commercial reserves exceed 43 Mt.

Diversity and a variable nature are typical for the ores, which is also manifested in variable concentrations of the main elements and of the silicon module value: Al₂O₃ - 34-76 %, SiO₂ - 1.5-21 %, Fe₂O₃ - 2-40 %, TiO₂ - 2-4.6 % and silicon module: 2.1-50 %. Accordingly, the mineral contents are also very variable. The main ore types are: haematite-boehmite, haematite-chamosite-boehmite, chamosite-boehmite, haematite-kaolinite-boehmite, kaolinite-boehmite and boehmite, of which the first two are most common; In the Central bed kaolinite-boehmite...

<table>
<thead>
<tr>
<th>Ore bed or field</th>
<th>Length, m.</th>
<th>Width, m.</th>
<th>Ore body thickness, m.</th>
<th>Depth of occurrence, m.</th>
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<td>1.5-21.0</td>
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<td>0.6-3.6</td>
<td>18.4-41.9</td>
</tr>
<tr>
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<td>100-400</td>
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<td>0.6-50.0</td>
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<tr>
<td>Low-iron bauxite area</td>
<td>460</td>
<td>100-400</td>
<td>0.6-2.4</td>
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</table>

Table 3: Parameters of the Vezhau-Vorykva deposit bauxite ores (Plyakin, 2005).
and boehmite types with a broad range of intermediate varieties are also important, many of which are suitable for abrasive production. The contents of trace elements in the ores of the Vezhayu-Vorykva deposit are normal for laterite-type bauxite, being (in g/t): Ga; 35-110; Sc, 40-140; V, 120-630; Nb, 30-90; REE, 600-1350. Since 1993, the deposit mining license has been owned by Boksit Timana JSC. The conditions allow open-pit mining; overburden rocks (mainly basalts) may be utilized as construction materials. Commercial development of the deposit began in February 1998; about 2 Mt of bauxite had been produced by 2002.

Upper Schugor bauxite deposit

The Upper Schugor deposit is located NW of the Vezhayu-Vorykva deposit. It consists of a South and North group of ore beds, which main difference is the different compositions of their decayed substrate. The North bauxite group was formed on feldspar-carbonate metasoamites; the ore bodies are heterogeneous, and are commonly split into a series of strata of various thicknesses, by alumina-rich laterites (allites) and clayey rocks (Fig. 46). The thickest ore strata (10 m and more) are confined to the top half of the ore body. The South bauxite group, formed on shale-carbonate rocks and meta-marls includes ore bodies usually consisting of one stratum occurring in the central part of the ore. The bauxites of the South group do not basically differ from the bauxite ores of the Vezhayu-Vorykva deposit in their chemistry and mineral composition, except for slightly higher contents of diasopore and aluminophosphate. A large part of the bauxites is usable for production of abrasives (5.7 Mt).

The North bauxite group contains more high-quality (high-module) grades, with haematite-boehmite, boehmite and kaolinite-boehmite mineralogy predominating. Very high silicon-module varieties (silicon module >28) varieties are frequent. Of special interest is a large unit of stained self-discoloured, iron-free high-module boehmite type bauxites. The average composition is (%): SiO₂ - 1.48; TiO₂ - 4.35; Al₂O₃ - 76.97; Fe₂O₃, 0.78; FeO - 0.06; MgO - 0.13; FeO - 0.06; Al₂O₃/SiO₂, 91.65. In kaolinite-boehmite bauxites, the average silica content increases to 10.66 %, and the silicon-module value drops to 9.77. High-module haematite-boehmite bauxites are also quite widespread in the North ore group, containing, on average, 0.88 % SiO₂, 61.26 % Al₂O₃ and 20.09 % Fe₂O₃ (silicon module: 69.6). Low contents of chamosite are a feature of all the bauxites in the Northern ore group. The main parameters of several of the bauxite beds of the Upper Schugor deposit are shown in Table 4.

The average composition of the Upper Schugor deposit is (% dry basis): SiO₂: 6.61, TiO₂: 2.87, Al₂O₃: 49.76, CaO: 0.39, S: 0.04, Ga: 0.0071, V₂O₅: 0.049; silicon-module value, 7.53.

The bauxites of the North ore group are characterised by extremely high concentrations of niobium and REE, which is due to REE mineralization in the parent metasomatic rocks. The average Nb contents are 200-700 g/t, in some samples up to 2000 g/t, present in pyrochlore, columbite, and niobotitanates. The bauxite sections show thin ore streaks with 2-5 % of Nb. Brown high-module bauxites contain 0.1 to 5.4 % of REE and up to 4.5 % Sr, present in monazite, apatite, or pyrochlore. These elements are not produced at the present.

Timsher (South Timan Bauxite District)

The Carboniferous bauxites in this district form the southern part of the Timan bauxite ore

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<th>Ore body thickness, m</th>
<th>Depth of occurrence, m.</th>
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<td>75-600</td>
<td>0.4-50.8</td>
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<td>0.6-47.6</td>
<td>12.64</td>
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</tbody>
</table>

Table 4: Parameters of the Upper Schugor deposit ore fields (Plyakin, 2005)
province. Stratigraphically, they are confined to the Visean stage, occurring on top of Devonian clayey carbonate sediments, and overlain most commonly by coal, and less frequently by Quaternary sediments. The South Timan bauxite district is mainly located 30-40 km, but up to 150 km, SSE of Ukhta. Two groups of deposits are identified within the district, Timsher-Puzla and Kedva-Volskaya.

The Timsher-Puzla deposits contain high-alumina, high-silicon, low-module, mostly pyritized (sulphurous) bauxites, characterised by a low iron content and white / light grey colour. The average depth of the bauxites is about 60 m; their thickness normally varies from 1.0 to 3.0-3.5 m, and, rarely, reaches 10 m. The bauxites of the Kedva-Volskaya group differ from those of the Timsher-Puzla group due to their low sulphur content. The Timsher deposit is located on the Vychegda-Izhma divide, along the R. Vychegda, furthest from Ukhta (120-150 km) compared to the other South Timan deposits. Structurally, it is confined to the south-western slope of the Timsher rise. Several beds are present in the deposit, the largest ones being the Western, Ezhvador, and Timsher beds. They are all near each other, forming one ore field. The First Timsher, Western, and Ezhvador beds have a sheet-like isometric shape with a wavy outline. The roofs and bases of the ore bodies are relatively uniform. White and grey kaolinite-boehmite bauxites predominate; red haematite-kaolinite-boehmite bauxites are less frequent. The following average contents of basic components are found in the bauxites (%): $\text{Al}_2\text{O}_3$ – 51.6; $\text{SiO}_2$ – 20.2; $\text{Fe}_2\text{O}_3$ – 5.44; $\text{S}_{\text{total}}$ – 1.84.

**Gremyakha-Vyrmes Fe-Ti deposit**

The Gremyakha-Vyrmes Fe-Ti deposit is located in the Kola District, Murmansk Oblast and is hosted by a layered ultramafic-mafic complex within a mainly alkaline-gabbro massif of Proterozoic age (1973 Ma, U-Pb age on monzodiorite, Vursii et al. (2000)). The massif is located in Mesoarchaean gneisses and granite gneisses. Its surface dimensions are 19 km N-S and 4-6 km E-W. The massif comprises four major complexes (from oldest to youngest): 1: ultramafic– mafic (marginal, gabbro wehrlite, and monzodiorite series and anorthosites); 2: alkaline complex (ijolites, melteigites, and foyaïtes); 3: complex of alkali granites and granosyenites; 4: complex of alkali metasomatites.

The complex Gremyakha-Vyrmes apatite-titanomagnetite-ilmenite deposit is related to a layered series of gabbro-anorthosite-peridotite rocks (Fig. 47). The ore deposit is hosted by the alkaline gabbro part of the layered intrusion. The most promising resources are titano magnetite-ilmenite ores found in the south-eastern part of the complex (Southern field), containing, on average, 10-20 % $\text{TiO}_2$. The overall calculated ore reserves in this part of the ore body are estimated to be 330 Mt (Category C1+C2), with an average content of 10.7 % $\text{TiO}_2$ and 23.17 % $\text{Fe}_{\text{total}}$. An
additional 150 Mt of richer ores with 12.5 % TiO₂ and 25.94 % of Fe_total have been identified in the central part of the ore body.

The average TiO₂ content in the ilmenite concentrate is 47 %, while the TiO₂ content in the titanomagnetite concentrate is 9 %, and the content of Fe(total) ~58.4 %. Based on its composition and physical properties, the ilmenite in the Gremyakha-Vyrmes deposit (i.e. the Southern field) gives a virtually pure mineral concentrate, and is easily decomposed in sulphuric acid, and a good raw material for the sulphuric acid process for production of titanium dioxide pigment.

A feasibility study by the company Giredmet has shown the viability of mining of the titanium ores of the Southern Gremyakha-Vyrmes deposit and their processing, using available capacities in the Olenegorsk and Monchegorsk mills. Based on prospecting and evaluation efforts, the overall reserves of titanium and phosphorus ores were estimated to 1.7 billion tons, with average contents of 2.84 % P₂O₅ and 5.53 % TiO₂. Concentrates of both apatite, with 39 % P₂O₅, and of ilmenite, with 49 % TiO₂, have been produced.

The Kovdor massif

The Kovdor massif of alkali/ultrabasic rocks and carbonatites is located on the Kola Peninsula, about 30 km from the Finnish border, in the basin of the R. Kovdora, a tributary of the R. Yona. The massif is a large intrusion, dated at 380-420 Ma, and emplaced into biotite-plagioclase- and granite gneisses (Fig. 48). The massif covers an area of about 41 km² and is pear-shaped in plan, with a zonal structure easily identified in the terrain. The ring structure is determined by long-lived injection of olivinite, ijolite-melteigite, and nepheline syenite, accompanied by silicate metasomatite and carbonatite.

The oldest intrusive rocks in the massif are olivinines forming its core of about 8 km² in horizontal section. The marginal zone of the massif is formed by alkaline rocks: melteigite, ijolite, and turjaite. Their injection along the contact of olivinites with the surrounding gneisses was accompanied by active alteration of both lithologies. Fenites are developed on the contact between the alkaline rocks and the enclosing gneisses of the White Sea series. The thickness of the fenite aureole varies from 0.5-1 km in the north to 2-3 km in the south of the massif. Mica-clinoxyroxene rocks occur between the olivinines in the core and the ijolites and turjaites replacing them.
The olivinites were enriched in mica and pyroxene during the ijolite stage, and partially replaced by monticellite, mellilite and mellilite-pyroxene metasomatites during the turjaite stage. Melilite-bearing rocks are commonly replaced by garnet-monticellite-phlogopite, garnet-monticellite-amphibole and garnet-diopside-amphibole (apomellilite) rocks.

A zone enriched in phlogopite forms a semi-circle 8 km long and 1.5-2 km thick along the periphery of the central olivinite core in the north. It consists of medium- to fine-grained phlogopite-diopside-forsterite rocks with lenses of large- to giant-grainsize phlogopite-diopside-forsterite metasomatites. Mining of phlogopite from the deposit, the largest in the world, started in 1965. The phlogopite rocks are cross-cut by feldspar ijolites. Forsterite, forsterite-apatite and calcite-tetraferriphlogopite lenses and veinlets are commonly developed in the diopside-phlogopite rocks. Baddeleyite and zirkelite occur in these lenses. Mellilite rocks in contact with rocks of the phlogopite complex were intensely transformed and replaced by hastingsite-calcite-diopside rocks. A monticellite-garnet association, called skarn-like by Kovdor massif researchers, is common in the mellilite and hastingsite-calcite-diopside rocks.

In the south-west of the massif, there is an ore body of forsterite-apatite-magnetite rocks (foscortes) and apatite-calcite-magnetite rocks. The ore body is located among ijolite, nepheline-clinoopyroxene and olivine-mica-pyroxene rocks. The ore body is crosscut by bodies of calcite carbonatites with green phlogopite, apatite, and magnetite, mainly on the south-west side, among fenites. Veins (up to a few metres thick) of carbonatite are also developed in the north of the massif cross-cutting ijolites, olivinites, mellilite rocks, mica-pyroxene rocks, and phlogopite rocks. Aegirine-calcite carbonatites with brown phlogopite and sphene occur in the south-eastern part of the massif. Dolomite carbonatites are developed exclusively within or in the immediate vicinity of the main ore body.

Formation of multiple carbonatite stockworks was one of the final stages in the evolution of the Kovdor massif. These rocks are highly diversified and of great commercial interest, because of the associated deposits of baddeleyite-apatite-magnetite, rare metals, carbonate and apatite-carbonate ores. Four stages of carbonatite formation have been identified:

1: Calcite carbonatites with forsterite, magnetite, phlogopite, baddeleyite.
2: Calcite carbonatites with forsterite (or clinohumite), magnetite, tetraferriphlogopite, baddeleyite, and uranium pyrochlore.
3: Dolomite and calcite-dolomite carbonatites in ijolites with ancolyte-(Ce), strontianite, catabrite, Nb-anatase and labuntsovite.
4: Dolomite carbonatites with tetraferriphlogopite, richterite, strontian whitlockite, strontian collinsite, girvasite, rinkorolgite, bobierite, krasnovite, and kovdorskite.

The youngest intrusive rocks of the Kovdor massif are nepheline- and cancrinite syenites, thin veins of which cut across all the above rocks, including the carbonatites. They are massive, grey, coarse-grained rocks, consisting mainly of feldspar, aegirine, diopside and nepheline. The nepheline is usually variably replaced by cancrinite, zeolites and other minerals. The ring-shaped bodies of alkaline rocks and mellilite-, mica-pyroxene- and other rocks situated between them and the olivinites dip at 70-80° towards the centre of the massif. The form of each type of alkaline rock and carbonatite is related to its own fault system inside the Kovdor massif (Krasnova & Sokolova, 1978).

A weathering crust is developed on top of the massif, covering ca. 60 % of its area. It is most widespread on the olivinites and phlogopite metasomatites of the central part, locally reaching a depth of 150 m. The formation of two commercial deposits is related to the weathering crust: vermiculite, by alteration of phlogopite metasomatites, and francolite by oxidation of the iron-ore complex rocks. The Kovdor massif contains five commercial deposits: a) Active mining of integrated baddeleyite-apatite-magnetite ore, phlogopite-vermiculite ore and rare earth elements b) Not in production: apatite-bearing carbonatites, olivinites, and francolite deposits. In addition, the kaolinite-lizardite ores confined to the weathering crust are also promising.

The Kovdor Iron Deposit

The Kovdor baddeleyite-apatite-magnetite deposit was discovered in 1933 simultaneously with the massif itself: the Kovdor mine and concentrating mill came into production in 1962. The iron-ore deposit is confined to the south-western part of the massif where it forms a vertically dipping tubular “Main body” with a cross-section of 800×1300 m and, in addition, some more linearly elongated bodies. There is clear evidence that the iron-ore complex was formed through magmatic replacement by carbonatites.

The deposit is divided into several ore types according to the content of characteristic minerals: apatite-forsterite, apatite-forsterite-phlogopite, apatite-forsterite-magnetite, apatite-calcite-magnetite, calcite-forsterite-magnetite, and apatite-calcite rocks. A less common ore type consists of magnetite, tetraferriphlogopite, calcite, and apatite. The ores vary in structure and may be banded, impregnated, spotty, and massive, but are generally characterised by a granular allomorphic texture. The grain size of the magnetite is from < 1 mm to several centimetres, with grains larger than 1 mm predominating. Common accessory minerals in all the different ore types include pyrrhotite, chalcopyrite, pyrite, marcasite, baddeleyite and pyrochlore (or uranium pyrochlore in rare cases). Sulphides are unevenly distributed. The magnetite has high contents of MgO (4.7-7.9 %) and Al₂O₃ (2-4.4 %). All parent rock species and carbonatites contain non-uniform weak, variable dissemination of baddeleyite. The ores contain: FeO 20-55 wt.% (average: 28.8 wt.%); MgO 15-17 wt.%; CaO 11-12 wt.%; P 2.9 wt.%; S 1.19 wt.%; MnO and TiO₂, < 1 wt.%.

The magnetite-apatite-baddeleyite deposit is being mined by Kovdor Mining and Concentrating Mill Co. using open-cast methods. The Kovdor Mill produces a magnetite concentrate with 64.0-64.2 % iron, a baddeleyite concentrate with 98.1-98.3 % ZrO₂ and an apatite concentrate with a minimum of 38 % P₂O₅. The ore reserves of the Kovdor deposit are 650 Mt.

The Kostomuksha Iron Deposit

The Kostomuksha deposit was discovered in 1946 as a result of a 1:200,000 scale airborne magnetic survey. It is located 12 km N of the city of Kostomuksha which was built to support the mining. The Kostomuksha mining and concentrating mill (now Karelsky Okatysh JSC) started full-scale ore extraction and pellet production in July 1982.

Conventionally, the deposit is subdivided into three units, the Northern, Central, and Southern units, differing in ore body shape and relationship with cross-cutting bodies of hälleflinta (microcrystalline to glassy felsic rock). Currently, all of the sites are being mined. The deposit is confined to horizons of iron quartzite which are part of the Kostomuksha Suite of the upper Lopian rhyodacite-iron-quartzite formation (Fig. 49). Up to 70 % of the reserves are in the main layer, which is situated in the western limb of a synclinal fold. The main layer consists of three steeply dipping, sheet-like ore bodies of iron quartzites. The bodies are 10 - 330 m thick and have been traced for 3.3 - 14 km in an approximately N-S direction. The ore bodies are separated by thin layers of schist, including quartz-biotite-sericite- and graphitic-schist. In the central part of the deposit, the Main layer is folded in such a way that it becomes sub-horizontal. The maximum width of the layer in the folded area reaches 1750 m while on the fold flanks, the width is from 13 m to 70-100 m. In the central part of the deposit, at a depth of 400 m, the thickness of the Main ore layer is 250-350 m, which decreases to 120 m at greater depths. The Main ore layer is 600 m long at its northern flank, 800 m at the southern flank, and 2100 m in the central part. Close to its extremities the ore quality drops because of increasing contents of grunerite.

The interbedded layers are situated 100-600 m E of the Main ore: they are characterised by rhythmic alternation of multiple (> 40) strata of iron quartzites and mica schists containing little or no iron. An ore reserve calculation considered 23 of the interbedded layers. These have strike lengths of 0.5 - 6.2 km, down-dip lengths of 100 - 500 m, and thicknesses of 5 - 130 m. With depth, some of the layers tend to increase in thickness, also having an improvement in ore quality. The ore zone has been traced for 16 km. It is intersected by drillholes...
at depths of up to 500-600 m on the flanks and up to 1000-1200 m in the central part of the deposit.

Three ore types are identifiable in the deposit: 1) Alkali-amphibole-magnetite quartzites containing 40-60 % magnetite, 30-50 % quartz, and <10 % alkali amphiboles (riebeckite, crossite, and aegirine). The ore has coarse-grained magnetite aggregates and is thus easier to process, 2) Biotite-magnetite quartzites with 30-50 % magnetite, <15 % biotite and locally up to 30 % carbonate (brown spar or magnesian dolomite), 3) Grunerite-hornblende-magnetite and grunerite-magnetite quartzites containing 35-50 % quartz, 20-35 % magnetite, up to 10 % pyrrhotite and up to 3% apatite. Sulphur and phosphorus are unwanted impurities. The average contents of magnetite iron (Fe$_{mag}$) decrease from the first to the third ore type. The predominant type in the Main ore layer is the first ore type, while the second (57 %) and third (22 %) types are most common in the interbedded layers. The third type is more common in smaller ore bodies and in the rims of large bodies.

<table>
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<th>Interbedded layer</th>
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<td>TiO$_2$</td>
<td>0.09</td>
<td>0.11</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>2.71</td>
<td>3.34</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>25.96</td>
<td>21.37</td>
</tr>
<tr>
<td>FeO</td>
<td>15.96</td>
<td>16.60</td>
</tr>
<tr>
<td>MnO</td>
<td>1.93</td>
<td>2.09</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>1.11</td>
<td>1.24</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.52</td>
<td>0.48</td>
</tr>
</tbody>
</table>

Table 5: Chemistry of the two ore layers:

The average contents for the whole deposit are: Fe$_{tot}$ 32.2 %; Fe$_{mag}$ 26.45 %; S, 0.21 %, P, 0.07 %. The ores are free-milling. Their dressing is a three-stage process of wet magnetic separation, producing a magnetite concentrate with an iron content of 65.7-70 %, while providing a recovery of Fe$_{tot}$ of 73.6-78.5 %, Fe$_{mag}$ recovery of 94.6-95.4 %, and concentrate yield of 33.8-37.3 %. The sulphur content in the concentrate is from trace levels to 1 %.

The commercial iron ore reserves were approved by the USSR State Committee for Reserves (Resolution No. 8668 of 19.12.1980) as prepared for mining. The quantity was 1,107,655,000 tons for Category B+C1 and 261,931,000 tons for Category C2. In addition, 1,023,025,000 tons of non-commercial iron ores were calculated, including ores that are too low-grade in terms of Fe$_{mag}$ content and ores beyond the open pit outline. With the mill's design capacity of 24 Mt of raw ore per year, the reserves were estimated to last for 45 years. The iron ore resources are currently calculated to be 300 Mt (Category P1).

Olenegorsk iron deposit

The Olenegorsk iron ore deposit was discovered by D. V. Shifrin in 1932. It is 7.5 km NW of the Olenya railway station in Murmansk Oblast. Open-pit mining started in 1954. The sheet-like ore layer extends 4 km NW, and dips 55-70° SW (Fig. 50). The ores have been explored to a depth of 800 m. The ore thickness varies from 20-30 metres at the flanks to 250-300 metres in the centre of the deposit. The ore body consists of magnetite and haematite quartzites with an average magnetite/haematite ratio of 3:1. The accessory minerals of the iron quartzite ore are tremolite, actinolite, cummingtonite, pyroxenes, alkali amphiboles, calcite, and siderite. Sulphides - pyrite, pyrrhotite and chalcopyrite are found but are rare. The ore is generally banded and “corrugated”. The deposit is characterised by extremely low contents of sulphur and phosphorus, (of the order of 10-100 ppm). Processing yields a very high ore mineral recovery to concentrate: 91 %. The ore is free-milling.

The iron quartzites of the Olenegorsk deposit are up to 150 m thick and extend for 2.8 km: they are hosted by amphibole-bearing gneisses and amphibolites of the Archaean Kola series. Gradual transitions between the iron quartzites and gneisses have been observed. In plan view, the iron quartzite layer has the shape of a large lens pinched by transverse flexures. The quartzite is deformed into a synclinal fold overturned south-westwards and complicated by anticlinal and synclinal folds of lower orders. The synclinal structure has a north-westerly strike, with the limbs dipping at 50-80° SW. The structure is intersected by low-angle, south-west- directed thrusts with amplitudes of up to 50-55 m. Flow cleavage, linear parallel textures and fracturing are pronounced features of the rocks in the
Figure 49: Schematic geological map of the Kostomuksha greenstone belt. (after V.Ya. Gor‘kovets and V.N. Furman. (1, 2) Dyke complex: (1) Riphean lamproite, (2) alkali picrite; (3) Late Archean–Early Proterozoic gabbro, unspecified; (4) Late Archean andesite and dacite dykes; (5–7) Late Archean late orogenic intrusive rocks: (5) potassium granite, (6) rhyodacite (helleflinta, plagiophyre), (7) Taloveis Complex of diorite, quartz diorite, and granite porphyry; (8) tonalite and granite gneiss; (9–11) Upper Archean Gimol Group: (9) schists of the Surlampi Formation, (10) quartz–biotite and amphibole–biotite schists and BIFs, (11) conglomerate; (12–16) Upper Archean Kontok Group: (12, 13) Shurlovaara Formation: (12) rhyodacitic lava and tuff, (13) carbonaceous quartz–biotite and amphibole–biotite schists, BIFs, and banded amphibolite; (14, 15) Ruvinaara Formation: (14) basalt and variolitic basalt, (15) komatiite, basaltic komatiite; (16) Niemijarvi Formation: metabasalt (amphibolite); (17) Nyukzero Sequence: two-mica schist; (18) fault; (19) iron ore; (20) gold deposits and occurrences (numerals in map): 1, Taloveis; 2, Faktorny; 3, Berendei; 4, Kurgelampi; 5, Eastern; 6, Niemijarvi; 7, Kostomuksha open pit; 8, South Kostomuksha; 9, West Ruvinaara.
The magnetite quartzite reserves in Categories A + B + C1 are 440 Mt, with an average iron content of 32.3%.

**Yarega titanium-oil deposit**

The Yarega titanium-heavy oil deposit is located in the Republic of Komi, 25 km SW of Ukhta in the Timan-Pechora oil and gas province. The deposit was discovered by the Ukhta Integrated Prospecting Expedition in 1932 and has been explored for titanium since 1958. The deposit is hosted by Middle Devonian sandstones and is confined to a wide, flat asymmetric anticlinal fold northwest of the Ukhta-Izhma ridge, on the north-eastern slope of the Timan antecline. The form of the crest of the antecline is influenced by the Yarega, South Yarega, Lyael and Vezhavozh local rises. The Upper and Middle Devonian sediments in the area are commercially oil-bearing, with reservoirs in quartz sandstones.

The unique feature of the deposit is that, apart from the large oil resources, it also contains high concentrations of leucoxene. The lower part of the titanium-bearing sandstone is of Eifelian/Early Givetian age, and the upper part is of Pashian age. The deposit is interpreted as a buried placer. The productive stratum is 30-100 m thick and discordantly overlies Riphean slates. It is divided into two lithological layers which contain three placer deposits: 1) A lower, sheet-like placer with an average thickness of 14.5-21.4 m which contains 11.2% TiO₂ on average; 2) A middle placer 0.4-13 m thick containing 3.0–10.4% of TiO₂; 3) An upper placer with an average thickness of about 3 m and a TiO₂ content ranging from a few percent up to, locally, 21.9%. The lower lithological horizon consists of coarse-grained quartz sandstones with siltstone and argillite layers, while the upper horizon, contains polymict conglomerates and consertal quartz sandstones containing up to 30% leucoxene (TiO₂: 58.5–71.9%; SiO₂: 20–37.8%). The deposit emerged as a result of erosion of the weathering crust of the Riphean slates. Later study of the titanium mineralization has revealed high contents of niobium and tantalum.

Yarega is the largest titanium deposit in Russia. The calculated reserves are about 640 Mt of titanium ore. The oil in the reservoir has been produced since 1939 by a unique shaft technique. In 2004, engineers of LUKOIL-Komi made some

![Figure 50: Geological structure of the Olenegorsk deposit (L. Goryainov, 1976)](image-url)
experiments to recover titanium dioxide and titanium coagulant from the Yarega ores. Work is currently underway to set up a pilot titanium coagulant production facility, to be managed by SITTEC JSC. The start-up of the first stage of a titanium coagulant mill for an annual output of 25,000 tons of finished product took place in the last quarter of 2015.

**Srednaya Padma vanadium deposit**

The Srednaya Padma vanadium deposit is located in the Medvezhegorsk District, Republic of Karelia, 17 km SW of Tolvui. The deposit is located on the Zaonezhsky Peninsula, which is a part of the Onega Trough, within the Karelian megablock (Fig. 51 and 52).

Exploration of the deposit area started with studies of shungites in the late 18th C. The vanadium deposit was discovered in 1985 as a result of prospecting by PGO Nevskgeologiya. In addition to Srednaya Padma, five other deposits and over ten promising occurrences of vanadium (with associated uranium, PGE, gold, silver, and other valuable components) were found – ores with unique mineralogical, geochemical, and technological properties. The Verkhnaya Padma
and Kosmozerskoye deposits are unique, with no known direct analogues in Russia or anywhere else.

The deposit is structurally confined to the south-western flank of the Padma anticline having, a symmetrical form and limbs dipping at >75°.

The richest mineralization is observed on the south-western flanks of the axial anticlines, in areas of intense multi-stage, tectonic processes of bulk brecciation and cataclasis. The ore bodies have a complex but generally cigar- or vein-like shape and a wedge-shaped cross-section. Less frequently, the ores occur in stockwork/veins traceable to a depth of 500-600 m. All the ore deposits have a polymineralic multi-element composition. The main recoverable elements are vanadium, uranium, palladium and gold, but a number of other elements may also be extracted, including rhenium, rhodium and others.

Vanadium is the main ore-forming element. The most common ore mineral assemblage is pitchblende-roscoelite-sulphide, which may contain precious metals. The ores consist of albite and carbonates (60 %) and mica (25 %). The bulk of the vanadium (90 %) is held in micas (mainly roscoelite, some in vanadium phlogopite), in vanadium-bearing haematite (10 %), and a lesser part in other oxides, e.g. uranium vanadates. The ores occur as impregnations (38 % of the explored reserves), massive mineralization (50 %) and in vein in stockwork (12 %). The ores are characterised by high contents of vanadium pentoxide, on average 2-5 %, but locally up to 15-17 %. The precious metal mineralization occurs in stockwork/veins and is usually developed within bodies of mica. The content of palladium reaches 140-440 g/t, of gold up to 120 g/t, of platinum up to 30 g/t and of silver up to 1500 g/t. Uranium occurs in all the ore deposits, usually as pitchblende and coffinite, less frequently as broggerite.

The average V_2O_5 content in the deposit is 2.78 % for Reserve Category C1, and 1.97 % for Category C2. The reserves of uranium-vanadium ores have been calculated to a depth of 350 m: Category C1 reserves of V_2O_5 are 58,770 t, and Category C2 reserves are 48,880 t.

Porozhinskoe manganese deposit
The Porozhinskoe manganese deposit was discovered in 1974, 650 km N of Krasnoyarsk, in the north-western part of the Yenisei Range, 12 km E of the R. Yenisei. It is located in the Mikheevskaya depression in the eastern part of the Vorogovka trough. The depression formed in Upper Riphean-Cambrian sediments deformed
by a complex system of linear-brachyform folds, later disrupted by NW- to NE-trending faults.

The sediments in the Vorogovka trough are divided into two series; the lower part is the Vorogovka series and the upper part the Chap series. The Vorogovka sediments form narrow bands on the western and eastern sides of the trough: they comprise conglomerates, gravelstones, sandstones, siltstones, limestones and dolomites. The total thickness of the series is 2.2-4.3 km. The series is subdivided into three suites which, from the bottom, are the Severnaya River, Mutnaya River and the Sukhaya River suites.

The Chap sediments overlie the Vorogovka series concordantly. The Chap sediments are subdivided into two parts. The lowermost, the Podyom Suite, contains dolomites, pyroclastic and terrigeneous formations. The upper part consists of terrigeneous rocks of the lower Nemchany Suite, consisting of red arkosic sandstones. The main manganese ores occur in the Podyom Suite, in persistent layers on weathered dolomites. The lower part, about 2 m thick, is represented by variegated manganiferous clays, in which fragments of dolomites or siltstones (aleurolites), and rarely of manganese ores occur in variable quantities. On top of this clay-rich layer there is a 4.3 m layer of grey brecciated silica rocks, followed by terrigeneous formations consisting of siltstones, argillites and thin layers of tuff, overlain by a sandstone unit. The siltstone and tuff rocks contain accumulations of manganese minerals. The productive layers are 22 - 85 m thick and occur at several stratigraphic levels, but mainly in the lower part of the Podyom Suite. Both primary and secondary ores are found in the deposit. The following types have been identified, based on their mineral composition: carbonatite or manganocalcite ores, semi-oxidized or todorokite-manganocalcite ores, and oxidized or pyrolusite ores. The secondary ores, mainly pyrolusite ores, predominate. The main ore minerals are pyrolusite, manganese, goethite, cryptomelane, bog ore, vernadite, manganocalcite, hydrogoethite, and todorokite. The ore textures are clayey-earthly-lumpy (nodular), lumpy-clayey-earthly and clayey. As of 2012, the ore reserves of the Porozhinskoe deposit were 15,696,000 t for Categories A+B+C1, and 13,767,000 t for Category C2, with an average manganese content of 18.85 %. The deposit development license is owned by Turukhansky Meridian LLC, now preparing for exploitation of the deposit.

Porozhinskoe bauxite deposit

The Porozhinskoe bauxite deposit is located in the Irkineev culmination of the Yenisei Range, confined to the Irkashevo anticinal rise as part of the sequence of Upper Proterozoic carbonate and shale rocks (Fig. 53). It contains several bauxite ore fields, such as Porozhinskoe, Artyugino and others. In each field, several ore deposits of the contact-karst type are known (over thirty in all): they are confined to karst depressions at the contact of shale rocks of the Krasnogorsk terrigenous Suite with carbonate rocks of the Potoskui series of the Upper Proterozoic Djura Suite. Some karst ore deposits are identified in karst depressions of the polje (interior valley) type and in individual small karst depressions confined to carbonate sediments of the Upper Proterozoic Aladnino Suite. The bauxite beds and ore bodies are located among variegated, brick-red kaolinite and hydargilitic-kaolinite clays. Their shape is complex, pocket-like in sink holes and lens-like in erosional karst depressions. The lateral extent of the ore bodies is 130—2000 m, and the depth of their occurrence, 0.1—94 m. Ore bodies with high-quality bauxites are confined directly to the zones of contact to the Krasnogorsk clayey shales with carbonate rocks. The bauxite quality deteriorates with distance from the possible sources of alumina for weathering. The average contents of the main components in the bauxites are 43.6 wt.% Al₂O₃, 13.1 wt.% SiO₂, 17.6 wt.% Fe₂O₃, 2.6 wt.% TiO₂, and silicon module (Al/Si) = 3.4. Basically, three types of bauxite ore are present: stony (11 %), loose (42.8 %) and clayey (38.4 %). The bauxites are of the gibbsite type. The primary ore-forming minerals are generally gibbsite, kaolinite, corundum, magnetite, haematite and goethite. Other minerals found are quartz, sphene and rutile. The mineralogy of the bauxite species is constant, but the quantity of the different minerals varies.

Pudozhgora Fe-Ti deposit

The Pudozhgora iron-titanium deposit is situated in the Pudozh District of the Republic of Karelia, on the east coast of Lake Onega, 6 km from Rimskoye settlement. The deposit was
Figure 53A: Schematic geological map of the Palaeozoic basement and bauxite-bearing sediments of the Porozhinskoe deposit. Based on Angara Expedition data. 1: Bauxite ores; 2: Cretaceous-Paleogenic kaolinite-bauxite-bearing sediments, kaolinite-gibbsite variegated clays with bauxite ores Cr—Pq; 3 – 15: Palaeozoic basement rocks: 3: Bituminous dolomite, dolomitic limestone, dolomite and argillite of L. Cambrian Lena age, Cm1I, 4: Sandstones with lenses of dolomites and argillites of Mashakovo suite Pt1 — Cm1mš, 5: Sandstone, argillite, dolomite of the Chistyakovo suite Pt1 — Cm1čs, 6: Sandstone, aleurolite, argillite of the Aleshkino suite Pt1 — Cm1ăs; 7 – 9: Sediments of the Potoskui suite Pt1pt: 7: Clayey shales, argillites of the Jura series Pt1pt2, 8: Carbonate rocks of the Jura series, 9: Clayey shales, argillites, sandstones of red series Pt1pt1; 10: Dolomites, limestones of the Aladnino suite Pt1al; 11: Clayey limestones, dolomitic limestones of suite of Card Pt1hrt; 12 – 14: Sediments of the Pogoryui suite: 12: Aleurolite-clayey and clayey shales, sandstones, quartzites Pt1 pg2, 13: Sandstones, quartzites, aleurolites Pt1 pg1, 14: Aleurolites, sandstones, aleurolite-clayey shales Pt1 pg1; 15: Clayey shales of the Udergei suite Pt1ud; 16: Tectonic faults.

Figure 53B: Geological section of the Porozhinskoe deposit. Based on the Angara Expedition data. 1: Quaternary formations; 2: Clayey, loose and stony bauxites; 3: Bauxite clays, alitites; 4: Variegated clays; 5: Clayey shale; 6: Limestones, dolomites, dolomitic limestones.
discovered in 1859 by Captain Anossov of the Mining Engineers Corps. The Pudozhgora Mining Company was established in 1898, but the deposit has not been mined to this day.

The main economic mineralization is titanium magnetite (Ti-V-Fe) ore: a Pd-Pt-Au mineralization, mainly associated with copper sulphides, is less important. The deposit is confined to an early Proterozoic poorly differentiated gabbro dolerite intrusion filling a near-horizontal fissure in Archean granitoids of the Vodlozero block (Fig. 54). The strike of the ore-bearing massif is north-westerly, and its dip is 30° - 48° SW. The intrusion has been traced for 7.1 km along strike and is 130-180 m thick in its central part, decreasing to 40-50 m towards its limits. The intrusion has a sheet-like, irregular shape with wavy contacts matching the fissure planes in the enclosing granites.

The titanium magnetite mineralization is a uniformly rich impregnation, occurring in three layers with strike lengths of 1000-3000 m, and 7.2 - 23.2 m thick (average: 14-17 m), occurring parallel to the footwall contact of the intrusion, on average 30 m above its base (in the upper, gabbro part of the section). The mineralization has been traced to a depth of 380 m. Two ore categories have been identified based on their contents of titaniferous magnetite, ores with 45 -75 % and 25 -45 % magnetite, respectively. Most of the ore layer falls into the richest category. The ores are persistent in strike and dip throughout the ore-bearing intrusion. Their thickness varies from 5.95 - 23.5 m, averaging 11-13 m. The economic components of the ore are iron, titanium, vanadium. Copper, gold, and platinum group metals are locally enriched in a sulphide-rich upper part of the Fe-Ti ore layer (3-8.5 m thick). The average contents of the ore are: Fe 28.91 %; Ti 8.13 %; V 0.43 %. The sulphide-rich layer contains on average: 0.13 % Cu, 0.21 g/t Au and up to 0.51 g/t Pt and 1.11 g/t Pd. The titaniferous magnetite ore reserves are 316.69 Mt but are currently classified as non-commercial. Development of the deposit is considered to be unprofitable due to the high energy costs of processing the titaniferous magnetite ore and to the complexity of the required metallurgical process.

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Deputatsky tin deposit

The Deputatsky tin deposit is located in Ust-Yana District, Republic of Saha-Yakutia, not far from the settlement of Deputatsky. The deposit was discovered by G. I. Kolmakov, member of the Irgichansk field crew of Dalstroy. The deposit was opened for production in 1951. The Western mine of the deposit was closed in 1999 and the operation was finally closed in 2009. The deposit is located in the western part of the Poloussny megasynclinorium marked by widespread flat dislocation blocks, which are delimited by discontinuous tectonic fractures of a fault-heave and thrust-heave nature, and by linear fold zones. The area consists of Jurassic terrigenous strata with an overall thickness of ca. 5 km, resting conformably on Upper Triassic sediments in the south (Fig. 55). The known tin ore deposits and occurrences are mainly located in flat dislocation blocks, where they are limited to the exocontact zones of small stocks of granite porphyry or granites, or to hornfels zones above hidden intrusions.

Only one small block of granitoid crops out on the surface in the ore district; another intrusion bulge was intersected at a depth of 377 m by a structural drillhole under the central part of the Deputatsky deposit. The area of the subsurface granite bulge within the Deputatsky ore cluster has been assessed by geophysical data to be ca. 50 km². Its depth is from 300 to 800-1200 m. Multiple indirect data indicate the presence of a large granite block, in an area of hundreds of km² at depth (Smirnov, 1978). K/Ar dating of biotite yielded ages for the granites of 143 million and 138 million years (Yablokov and Ivanov), which corresponds to Upper Jurassic (bulk sample tests yielded slightly lower and probably underestimated values, corresponding to Lower Cretaceous).

Apart from granites, the sedimentary block is intruded by multiple mafic, intermediate, and silicic dykes. These include dolerites, doleritic and diorite porphyrites, andesitic dacites, liparites, felsitic liparites, and lamprophyres. The dykes of the more mafic rocks form E-W striking suites, up to 15 km long. Felsic rocks are less common and typically strike in near E-W or north-easterly directions. Assessments of the dykes' absolute ages have given a broad scatter, from 128 million to 70 million years, but mostly indicate Upper Cretaceous ages. Many of the dykes of dolerite, porphyrite, and especially of lamprophyre contain xenoliths of granite porphyry, quartzite, and metamorphic rocks carried up from the deep.

The Deputatsky deposit is located on the flat-lying limb of a large fold, dipping towards a hidden bulge of a granite body. The Upper Jurassic stratum of sandstones with subordinate siltstones form the ore field, dipping at angles of 4—20°S (rarely up to 30° or more) where the folded structure is cut by an E-W striking fault. In the central part of the deposit, the sedimentary rocks underwent intense contact metamorphism, generating mainly two-mica and andalusite-bearing biotite-quartz hornfelses. Quartz-muscovite, quartz-topaz and quartz-tourmaline greisens with tin ore, arsenopyrite and fluorite formed over offshoot dykes of granite and in hornfelses in zones of jointing. The intensely metamorphosed area is surrounded by a zone of less modified rocks, which are nevertheless noticeably silicified, chloritized and sericitized, and locally have a nodular texture with new formation of andalusite, cordierite, and axinite. Such rocks cover up to 65 % of the entire ore field and are most abundant at the western edge of the deposit. The ore field’s peripheral part is marked by weak silicification, sericitization, and local pyritization.
The ore field boundaries approximately coincide with the boundaries of weakly metamorphosed rocks. They extend in near E-W direction concordantly with the strike of sedimentary rocks, main dyke suites and principal ore faults.

In all, there are about 150 ore bodies in the deposit. They are classified, according to their morphological features, into three types: veins, linear elongated stockwork-like zones, and mineralized cataclastic zones traceable for many hundreds of meters, with thicknesses of up to 10 m and more; combinations of two or all three types are common. Most of the ore bodies are related to thick, extended mineralized shear zones, within which a central fissure vein generally shows up as the most persistent one laterally and down-plunge, accompanied by a series of parallel apophyses, zones of crushed and mineralized rocks, en echelon fractures of tear and shear, also consisting of ore material.

The mineralized cataclastic zones are marked by a complex morphology, diversified mineral composition, and rather uniform, high content of tin; their strike is from north-west to north-easterly, plunging 75-85° S. Vein-type ore bodies, with a similar steep southward plunge, with east-west and north-east strike, are much thinner and have relatively limited lateral extent on the order of a few hundreds of meters; forking of veins and their transition from one fracture set to another is observed. Stockwork-like zones typical for the deposit’s eastern parts also are not extensive, but their width is measured in tens of metres.

Primary minerals predominant in the ores are quartz, pyrrhotite and, locally, tourmaline, chlorite and pyrite, typical mainly for altered wallrock. Ubiquitous secondary minerals include cassiterite, chalcopyrite, sphalerite, manganosite, siderite, and ankerite. Marcasite, fluorite, and calcite occur locally in large quantities. Rare minerals include arsenopyrite, galena, sericite, topaz, wolframite, albite, boulangerite, jamesonite, frankeite, proustite, pyrargyrite, falsores and bismuthinite. The most widespread...
hypercene minerals are limonite, jarosite, fibroferrite, melanterite, scorodite, pisantite, melnikovite, kaolinite, and gypsum. The host rocks have been hydrotermally altered, most commonly silicification, tourmalinization and chloritization and less intensely sericitization, sulphidization, and carbonatization. The ore textures include brecciated, massive, and ribbon textures.

Four main types of ore, in terms of mineralogy, are found: 1) quartz-tourmaline veins with cassiterite; 2) cassiterite-sulphide-quartz veins with tourmaline and fluorite; 3) cassiterite-chlorite-sulphide mineralized cataclastic zones and veins; 4) quartz-carbonate veins with sphalerite and galena; weakly tin-bearing greisen formations are also found. Ores of the second and third types formed in several stages have the largest economic importance.

The predominant part of the Deputatsky deposit’s resources is in oxidized ores containing relics of primary ores. The total depth of the oxidation zone in the deposit is 400 m and more. The upper part of the oxidation zone consists of jarosite-limonite ores. The deposit’s tin reserves, as of 01.01.2014, were: Proven (Categories A+B+C₁), 198,300 tons; Inferred (C₂), 57,500 tons, with an average tin content in the ores of 1.15 %. The reserves show that it is the largest tin deposit in Russia. As of 2014, the license for its development was held by Deputatsky Mining & Concentration Works JSC. There has been no ore production at the deposit in recent years.

**Odinokoye tin deposit**

The Odinokoye tin deposit was discovered by M. I. Ipatov in 1945. It is located in the southern part of the Polous synclinorium, on the eastern flank of the Deputatsky brachysynclinal zone. The deposit is located in a granite porphyry stock penetrating the northern limb of the Odinokoye E-W trending brachysynclinal fold structure re-folded in an anticlinal structure. The structure consists of Upper Jurassic, predominantly sandstones with lesser shale, locally metamorphosed by the Omchikandya intrusion.

The granite stock is confined by the intersection of NE- and NW-striking faults, the faulting separating the brachysynclinal system from the district of near E-W trending linear folds of the Bakyn anticline to the north. The outcrop of the stock has an ellipsoidal outline, extending in a northeasterly direction and complicated by many short apophyses in a northwesterly direction, as well as elongate apophyses in a northeasterly direction. At the northern contact, the intrusion consists of eruptive breccias with fragments of both country rocks and the granite porphyries.

The porphyry rocks are intensely greisenized, except for the central part of the outcrop where the granite porphyries are only weakly modified by autometamorphism. Their chemical composition indicates that the intrusion is a highly siliceous granite with excess aluminum, potassium predominating over sodium, and a high content of alkalies, and trace elements as follows (in g/t): lithium, 190; boron, 11; and tin, 273. Together with a series of dykes crosscutting the biotite granites, these rocks belong to a late phase of the granodiorite-granite complex.

The deposit forms an ore shoot in a typical roof position with an apophysis/inclusion nature of the ore body. The ore shoot was formed in the greisen zone, which preceded the ore productive phase. The extent of greisenization correlates with the position of the southern contact of the intrusion and faults which strike ENE and NW. Topaz-quartz greisens were formed along the stock’s granite porphyries, its apophyses and along the exocontact hornfels. A second facies is represented by topaz-mica-quartz, and a third by kaolinite-mica-quartz greisens, the latter being found only in the granite porphyry stock. The ore body is marked by a south-western pitch on the eastern flank: it flattens out in the central dome and takes an opposite westerly pitch in accordance with the form of the contact on the western flank. The tin content in the topaz-quartz and topaz-mica-quartz greisens increases by one order of magnitude as compared to the weakly auto-metamorphosed granite porphyries.

The Odinokoye deposit belongs to the greisen type of tin ore-quartz formation, with extensive development of micaceous iron ores. The ores are strongly enriched in fluorine-bearing minerals, including fluorite or topaz, and high iron contents in mica and other minerals. The major minerals in the ores are quartz, topaz and siderophyllite. Muscovite, kaolinite, fluorite, haematite, goethite, and cassiterite are less common and accessories include native bismuth,
sphalerite, pyrrhotite, chalcopyrite, molybdenite, lollingite, pyrite, marcasite, arsenopyrite, siderite, wolframite, apatite, chlorite. Galena, stannite, bornite, valleriite, cubanite, bismuthite, rutile, magnetite, columbite, tapiolite, samarskite, pyrochlore, gypsum, scheelite, monazite, xenotime, zircon, cordierite, tourmaline, and epidote occur locally. Hypogene minerals are represented by hydrogoethite, kaolinite, scorodite, psilomelane, azurite, malachite, and pharmacosiderite. Reserves of Category A+B+C1: 125,800 tons, C2 – 1,800 t. The deposit was developed from 1987 to 1993, but the operation was stopped and the deposit abandoned in 1994.

Pervonachalnoye tin deposit

The Pervonachalnoye tin deposit belongs to the Pyrkakai tin ore cluster located in the Chaun District of the Chukotka Autonomous Okrug on the east coast of the Chaun Bay 85 km from the town of Pevek, which is located on the Arctic coast at 170°E. The deposit was discovered by B. N. Yerofeev in 1938 during a geological survey by the Mleluveem crew of the All-Union Arctic Institute. The deposit was explored in the period 1964-1980 (trenches, drillholes and adits) and since then has been included in the State Reserve.

The Pyrkakai tin-bearing cluster is a part of the Kuiviveemo-Pyrkakai tin-bearing district covering an area of 4,000-5,000 km² on the east coast of the Chaun Bay, in a zone affected by the deep-seated, arc-shaped Palyavaam-Yanranai fracture. The district is marked by the widespread occurrence of early Cretaceous stanniferous granitoid intrusions. The known ore fields and occurrences are related to intrusive roof bulges, and in some cases are located directly within the bodies of eruptive massifs. The Pyrkakai cluster is located in the above-dome part of the “blind” East Chaun granite batholith which has been localized by geophysical surveys. The explored stockwork zones are related to local rises in that block, and are controlled by nodes of intersection of pre-mineral N-S-trending fracture zones with discontinuous northwesterly and near-E-W-trending faults.

The ore bodies of the Pervonachalnoye deposit are typical representatives of the stockwork morphological type (Fig. 56). The Krutoi, and Central stockworks are spaced 750-800 m apart, and several smaller ones - Feathering, Southern, and Eastern stocks, have been studied in detail. Other stock-work type occurrences are also known, as well as linear veined zones, mineralized cataclastic zones, and mineralized lamprophyre dykes. The surrounding rocks, mainly clayey shales with siltstone and sandstone lenses are subdivided into four layers 50 - 250 m thick (upward): siltstone-shale, shale, sandstone, and sandstone-shale. The first two layers are classified, on the basis of their fauna as being of Carnian (Late Triassic) age, and the others, of the younger Norian stage. Intrusives are represented by lamprophyre dykes, mostly with a N-S strike and less common granodiorite porphyry dykes. All the dykes are affected, to some extent, by metasomatic processes. According to the results of gravity and magnetic surveys, the ore field is located above the roof of the granitoid block at its transition from a horizontal to steeply dipping attitude, at a depth of 1.6-2 km to the granitoid block.

The sedimentary rocks hosting the deposit were hydrothermally modified as a result of tourmalinization, silicification, sericitization, sulphidization, and chloritization. Scattered occurrences of biotitized rocks have been recorded. Evidently, the hydrothermal alteration occurred mainly in a pre-mineralization phase. More local fields of hydrothermally altered rocks are related to wall-rock metasomatism. The fields of hydrothermally altered rocks have no clear boundaries and are usually spatially linked, forming quartz-sericite, tourmaline-sericite, and tourmaline-quartz rocks with transitional varieties between them. Sulphides are inherent to all the rock varieties.

A NW-trending anticlinal fold is present within the deposit, complicated by high-order antilines and synclines. Carnian sediments crop out in the fold core. The main role in forming the structural framework of the deposit and stockwork shoots is related to N-S, NW, and E-W-trending faults. All of the deposit’s stockworks are localized in the joint nodes of northwesterly-trending and E-W fractures in the Olenyino fault zone with concentration of small shear fissures, forming part of the N-S-trending Pervonachalnaya-Nagornaya fracture zone. The intensity of the mineralization in each stockwork decreases with distance from the Olenyino zone. The stockworks, i.e. the main ore-bearing structures, are linear vein systems. Apophyses are spaced at 20-25 cm across the stockwork and strike N-S with a deviation of ± 15° and a steep plunge. Two minor N-S-trending
fracture systems are commonly developed in each stockwork, with opposite plunges - westwards and eastwards at angles of 70-85°. The second system is generally represented by thin joints without mineralization, and locally consisting of pre-mineral quartz or small inclusions of early sulphides. The apophyses, which make up the productive ore, contain over 60 hypogene and hypogene minerals. Quartz (> 20 %) is the main mineral: muscovite, fluorite, topaz, pyrrhotite, arsenopyrite, pyrite, sphalerite, albite and chlorite are less common (1 - 20 %); rutile, apatite, tourmaline, siderite, calcite, wolframite, galena and cassiterite are rare (< 1 %). Inclusions of cassiterite reach a size of 15-20 mm, with an average of 5-6 mm. The large size of the cassiterite segregations, the absence of intergrowths with other minerals, and the negligible content of tin sulphide compounds permit categorization of the Pyrkakai stockwork ores as free-milling. The average tin content in the ore is 0.23 %. The reserves of the Pervonachalnoye deposit in Category A+B+C₁ are 243,000 t, and in Category C₂, 7,000 t.

The Svetloje tin-tungsten deposit

The Svetloje tin-tungsten deposit is located in the north of Central Chukotka, in the Amguema tungsten ore district, at the SE end of the Kuveemkai anticlinorium. The mineralization is related to granites of the Cretaceous Iultin type. The Svetloje deposit is located in the Northern tin-tungsten cluster. The rocks forming the ore field are Triassic sediments, primarily silt- and sandstone. The host rocks are contact metamorphosed by a granite intrusion outcropping at Mount Veshkap. The resulting hornfelses contain pyroxene-hornblende, mica-andalusite and tourmaline-mica but also include spotty and nodular schists and hornfelsed sandstones. The area is cut by faults with a north-easterly strike which control the distribution of the tungsten mineralization and of ore-free quartz veins. Post-mineralization tectonic faults are evident in the ore field. The displacement on these is up to 8.5 m. Various dykes occur widely within the ore field, confined to a north-westerly-striking deformation zone, the thickness of which is ca. 400 m and the length >1 km. Dykes...
of coarse-grained granite porphyry occur in the north-west, micro-granites in the central part, and aplites in the south-east. The dykes have a north-westerly strike (290 - 330°) and dip at 60° SW; their thickness reaches 50 m.

Two morphological types of ore bodies are identifiable within the deposit: 1) vein-type, developed mostly in hornfelsed rocks where ore bodies occur without any important alteration of the wall rocks, and 2) mineralized ore zones generated mainly in granitoid dykes subjected to intense alteration. The ore bodies of the vein type have higher contents of cassiterite and wolframite. The veins occur in sedimentary rocks and are persistent in strike but not in thickness and dip; locally they cross-cut dyke formations. The vein thicknesses vary within a broad range. Ore body thicknesses decrease sharply when the veins extend from sedimentary rocks into dykes. The veins have a north-westerly strike (300—310°) and dip 70-85°SW. The contacts with enclosing sandy shales are markedly linear; but in dykes, they are diffuse. At the contact with veins, a marginal zone of 3-5 mm of mica appears in the sedimentary rocks: intense greisenization is recorded in dykes (in zones up to 2 m wide).

The mineralized zones are not persistent in strike and dip. Thickened mineralized portions are replaced by thread-like veinlets within a few tens of metres. Sampling data indicate that large accumulations of ore minerals in such zones are confined to quartz veinlets and vein accumulation areas. At the contact with these zones, the granitoids are more intensely greisenized and contain a higher grade of tin mineralization.

The following mineralogically defined ore types containing wolframite and cassiterite are found: quartz, quartz-topaz, and occasionally quartz-arsenopyrite. The quartz type is predominant. The ore bodies have a varied mineralogical composition. Quartz is the basic mineral of the ore veins together with topaz, arsenopyrite, and wolframite. Muscovite, feldspar, cassiterite and pyrite are subordinate and fluorite, biotite, chalcopyrite, pyrrhotite, and löllingite are accessory: carbonates, scheelite, bismuthite, molybdenite, apatite, anatase, and sphalerite are very rare.

Wolframite occasionally forms large pockets up to 0.4 cm in size, associated with cassiterite ore. In some ore bodies, a noticeable increase in the concentration of wolframite with depth has been observed. The ratio of wolframite and cassiterite in the ore bodies varies from 2:7:1 in the vein type to 1:1-1:1.5 in mineralized zones. Wolframite is rather irregularly distributed throughout the ore body and occurs both in the fringes and central parts of veins. Cassiterite is found in greisens, quartz and arsenopyrite-quartz veins, and in the micaceous margins of quartz veins occurring in sandy shales. The highest tin contents occur in mineralized zones. Cassiterite is mainly associated with quartz, occasionally with topaz. It is mostly coarse-grained, and its distribution throughout the ore body is rather erratic.

Tomtor REE deposit

The Tomtor REE deposit is located in the north-west of the Republic of Sakha (Yakutia), within Olenek Ulus, on the divide of the Udja and Chymara rivers, 400 km S of the Laptev Sea coast. The alkali-ultramafic-hosted ore was discovered by E. N. Erlikh in 1959 during a 1:200,000 scale prospecting survey. Several areas with radioactive REE mineralization were discovered within the region in 1960. In the 1970s and 1980s, significant accumulations of iron, phosphorus and rare earth elements were found in the carbonatites and their weathered zone in the core of the Tomtor massif. The remote location of the deposit, despite its large resources, has delayed its development for several decades.

The Tomtor deposit is located on the western slope of the Udja rise, and confined to an alkaline, ultramafic and carbonatite massif of the same name (Fig. 57). The surrounding rocks are dolomites, shales, and argillites of the Riphean Ulakhan-Kurung suite and terrigenous metamorphosed rocks of the Vendian Tomtor suite. In plan, the massif has a round, nearly isometric shape ca. 20 km in diameter, and a total area of ca. 250 km². The structure is concentrically zoned. The central part, 4 - 5 km in diameter, consists of carbonatite complex rocks, which are a substrate for an ore-bearing, weathered mantle. The carbonatite core is flanked by ultramafic and foiditic rocks on its eastern and western sides.

The ore body is confined to a kaolinite-crandallite horizon of a re-deposited weathering mantle, occurs on siderite horizon ores in most cases.
and is overlaid by sedimentary formations of the Permian, Jurassic, and Quaternary age. The total thickness is 7.5 m - 160 m. The ore stratum is an alternation of layers variably enriched in pyrochlore and monazite with lean crandallite (Ba phosphate) and kaolinite-crandallite layers containing 1-2 % of niobium oxides and up to 3-5 % of rare earth oxides. No zoning has been found in the distribution of layers, but pyrochlore streaks are restricted to the core and base of the section.

The ores within the Tomtor ore deposit are subdivided into the following types according to their mineral and chemical composition and geological/genetic features:

- Ores in the re-deposited (epigenetically altered) weathering mantle (main economic type);
- Ores in bedrock carbonatites;
- Ores in sedimentary Permian deposits.
The economic ore reserves of niobium ore calculated for the Buranny ore field are 42.7 Mt. The contents in the ores are: Nb$_2$O$_5$, 6.71%; Y$_2$O$_3$, 0.595%; Sc$_2$O$_3$, 0.048%; REE$_2$O$_3$, 9.53%. Tomtor is also one of the world’s largest REE deposits. The niobium content per ton of rock varies from 23 to 63 kg, and for REE, up to 93 kg. The rights for exploration and mining of ores of niobium, REE, scandium, and associated components of the Buranny field of the Tomtor deposit were granted to Vostok Engineering LLC (subsidiary of TriArc Mining) in May 2014.

**Lovozero Nb-Ta-REE-rich Massif**

The Lovozero alkali massif was first described by W. Ramsay in 1887. The massif’s surface area is 650 km$^2$ and it is the world’s largest augpative intrusion. The massif is located in the central part of the Kola Peninsula, between Lovozero and Umbzero. It is of Paleozoic age and intruded into Archaean granitic gneisses and, to a lesser extent, Proterozoic sillimanite schists and jaspilites, and weakly metamorphosed Lower and Middle Devonian sedimentary and igneous rocks (Fig. 58). The massif’s rocks were formed in several intrusive phases. Rocks of the earlier phases and relics of the Lovozero stage have only survived in xenoliths inside the massif and in the preserved part of its roof and margins.

The Lovozero massif is related to a late Devonian province of alkali- and nepheline syenites with carbonatite. It was formed by three successive phases (Kalinkin, 1974). The first phase is represented by a stratified series of loparite-bearing lujavrite/foyaite/urtite, nepheline and cancrinite-nepheline syenites, with alkali syenites in the marginal zone. The second phase forms in several intrusive phases. Rocks of the earlier phases and relics of the Lovozero stage have only survived in xenoliths inside the massif and in the preserved part of its roof and margins.

The Lovozero alkali massif in the Kola Peninsula is one of the largest known stratified intrusions, related to which there are immense deposits of loparite ((Ce,Na,Ca)(Ti,Nb)O$_3$). The massif consists (Fig. 58) of a complex of eudialyte-bearing lujavrite building up the top part of the massif with alternating horizons of leuocratic, mesocratic and melanocratic varieties, as well as eudialyte-bearing lujavrite, foyaite and urtite. Underneath this there is a thick differentiated complex, formed by recurrent three-fold layers of foyaite-urtite-lujavrites. A complex of nephelinite, nepheline-hydrosodalite and poikilite syenites is concentrated mainly in the massif’s marginal parts and probably underlying the differentiated complex rocks, and a complex of veined alkali rocks (Bussen, Sakharov, 1967; Vlasov et al., 1959; Gerasimovsky et al., 1966). Loparite mineralization is mainly confined to the horizons of urtite, malignite, and less commonly lujavrite of the differentiated complex. A consistent change in the composition of the loparite has been identified throughout the massif (Vlasov et al., 1959; Ifantopulo, Osokin, 1979; Kogarko et al., 1996, and others). In a vertical section of the intrusion, the SrO, Nb$_2$O$_5$, Ta$_2$O$_5$, ThO$_2$, Na$_2$O contents increase upwards, but the contents of CaO, FeO, TiO$_2$, CeO$_2$, La$_2$O$_3$, Nd$_2$O$_3$, and total REE decrease. The loparite of eudialyte-bearing lujavrites and some pegmatites show an increase in lueshite (NaNbO$_3$) and tasonite (SrTiO$_3$) components compared to the loparites of the differentiated complex (Kogarko et al., 1996).
The Lovozero deposit comprises 12 ore fields, of which those currently developed (Karnasurt, Kedykvypakhk and Umbozero) account for 75% of the reserves. Other fields are earmarked for development in the foreseeable future. The operating underground mines have reserves for at least 55-100 years. The reserve base consists of eudialyte-loparite ore with the advantages of high contents of tantalum (up to 1%) and niobium (up to 10-12%). Additional reserves with other elements include deposits of eudialyte-loparite and eudialyte ore containing zirconium and yttrium in the Alluaiv field, which can be developed underground. The metal oxide content in loparite exceeds 80-85% (Ta₂O₅, 0.6%; Nb₂O₅, 7-8%; SrO, 0.7%; TiO₂, 40%; REE (oxide), 36%). The Lovozero mining and concentration mill comprises an ore dressing mill with an annual capacity of up to 1.5 Mt of ore. The dressing mills use a gravity pattern to produce 95% loparite concentrate, which is processed in the Solikamsk Magnesium Mill using chlorine technology to produce niobium, tantalum, and rare earths.

“Karnasurt” was opened in 1951 and “Umbozero,” in 1984. However, the Umbozero mine suffered a heavy man-made earthquake in 1999, as a result of which, the mine was closed (Kozyrev et al., 2002; Lovchikov, Savchenko, 2013). The Karnasurt mine, about 10 km away from the Umbozero mine, continues to be operated.

Kolmozero Li-Be-Ta-Nb deposit

The Kolmozero (A. E. Fersman) deposit of rare metal pegmatites was discovered by A. A. Chumakov and I. V. Ginzburg, researchers of the Kola branch, USSR Academy of Sciences, in 1947. The deposit is located in the Lovozero District, Murmansk Oblast, 80 km E of Lovozero settlement, in an unpopulated and undeveloped area. The Kolmozero deposit is located in the SE part of the Uraguba-Kolmozero ore belt, being part of
the Kolmozero-Voronya ore district, Iokanga ore field.

The deposit consists of 12 veins of albite-spodumene pegmatites localized in metagabbro-anorthosites of the Potchemvarek massif which is at the junction of the Kolmozero-Voronya greenstone belt and the Murmansk block. In plan, the metagabbro-anorthosite massif has a lens-like shape with a northwesterly strike (300–310°). Its length is ca. 7 km, and its maximum width, 2 km. The massif consists of metagabbro, meta-anorthosite and amphibolites, dipping steeply to the NE. The age of the metagabbro-anorthosite is 2925 ± 7 Ma (U-Pb zircon dating). The massif is cut by faults striking NW, approximately N-S and NE.

The metagabbro-anorthosite massif is, to the south, in contact with the Kolmozero-Voronya greenstone complex (dated at 2.92–2.65 Ga) and to the north with Murmansk block rocks, sanukitoids of the Kolmozero massif dated to 2736 ± 11 Ma (U-Pb zircon). The TDM model Sm-Nd age of the rocks is 2.9 Ga. A new Sm-Nd (TDM) model age obtained for migmatized granite gneisses of the Murmansk block is 3.1 Ga. In the contact zone, the rocks have been strongly shearedto form banded mylonites and ultramylonites, and later folded into asymmetric S-shaped folds. The contact of the metagabbro-anorthosite massif with migmatized granite gneisses is complicated by a zone of quartz-chlorite schists alternating with holmquistite schists, which result from hydrothermal alteration of the metagabbro-anorthosites during pegmatite formation. Alteration products are confined to a NW striking tectonic zone favourable for circulation of post-magmatic solutions.

Early pegmatite bodies, located near a granite source, are of limited length and thickness, and do not contain rare-element bearing minerals. Later, less viscous pegmatite melts were enriched in volatile components (H, F, Cl, P, S, B) and lithophile rare elements (Li, Cs, Sn, Rb, Ta). The pegmatite veins dip steeply to the SW, have a length of up to 1400 m and a thickness of up to 25 m; the strike is northwesterly with a slight deviation to the SW. Large veins are complicated by apophyses, swells, and pinches. The main structural elements controlling the locations of rare metal pegmatite veins are NW striking fractures with a dip of 50–70°. The rare metal pegmatites contain xenoliths of metamorphosed, foliated host rocks, which indicates injection of pegmatite melt into deformed and metamorphosed gabbro-anorthosites. The contacts of the pegmatites with the metagabbro-anorthosite wallrock are well-defined, and, in some cases, tectonized. Fine-grained acicular holmquistite (Li,Mg,Al,Si$_4$O$_8$(OH)$_4$) and biotite has been observed in the endcontact zone. The rare metal pegmatites are deformed and disrupted by fractures. The age of the pegmatites is 1994 ± 5 Ma (U-Pb zircon dating, L. N. Morozova, 2014).

Quartz-feldspar (ore-free), quartz-muscovite-feldspar (ore-free with beryllium) and albite-spodumene (ore) pegmatites are found within and around the Kolmozero deposit. Feldspar pegmatites are developed in all the above-mentioned rocks, but albite-spodumene pegmatites occur only in the amphibolites and metagabbro-anorthosites, and muscovite-feldspar pegmatites among the rocks of the Kolmozero-Voronya greenstone complex.

The deposit’s economic reserves in Categories B+C+D are 844,200 t of lithium oxide with an average content of 1.14 %. The average contents of tantalum, niobium, and beryllium (0.009 % Ta$_2$O$_5$, 0.011 % Nb$_2$O$_5$ and 0.037 % BeO) permit considering these components merely as accessory to spodumene production and their reserves are included in the register.

**Polmostundra Li-Ta-Nb deposit**

The Polmostundra rare metal deposit is located in the Lovozero District, 48 km NE of Lovozero settlement, in an undeveloped area. The deposit was investigated in detail in 1957-1960. Rare metal pegmatites were intersected by trenches at 50 m intervals and were explored in depth by densely spaced drilling.

The deposit comprises 12 en-echelon arranged veins of rare metal pegmatite with a northwesterly strike and a dip of 50°NE within an area of 3000 x 200 m. The host rocks are schistose amphibolites. The veins are 1000-1300 m long, 3 to 15-30 m thick, and have been traced down-dip for 200-300 m without signs of thinning. The pegmatites consist of quartz (30-40 %), albite (10-70 %), spodumene (20-50 %) and microcline.
(up to 20%). The average content of Li₂O is 1.25 %, and of its accessory components: Ta₂O₅, 0.004 %; Nb₂O₅, 0.007 % and BeO, 0.027 %. A block of rich spodumene ores with 2 % Li₂O (4.5 Mt) has been identified in the deposit. Dressing technology has been studied at a pilot production level, leading to concentrates of: spodumene with 5.1 % Li₂O (90 % recovery), tantalite-columbite of 8.7 % Ta₂O₅ and 30.4 % Nb₂O₅ with 28 and 59 % recovery, respectively; a beryllium concentrate has not been obtained.

Odinokoye placer tin deposit

The Odinokoye placer tin deposit is located on the border between the Ust-Yana Ulus District and Allaikha Ulus District, in the Republic of Sakha (Yakutia), along the boundary between the Kuranakh and Berelekh river systems. The deposit has no permanent communication routes, except for the Odinokaya – Omchikandya – Deputatsky – Ust-Kuiga winter motor road. The first information on the geological aspects of the deposit area dates from the late 1930s. In 1945, a geological reconnaissance revealed an ore body (Polyarnoye), a placer tin deposit (Omchikandya) and the tin occurrences of Mount Odinokaya. The Odinokoye tin ore deposit was, in the period 1970-83, the object of prospecting. The deposit was mined from 1987 to 1993. The operation was suspended in 1994 and the deposit was put under care and maintenance.

The Odinokoye tin placer deposit is a heterogeneous formation and includes the placer tins of the Odinoky, Mokry and Yasny streams and the slope placer of Mount Odinokaya. The deposit was formed by a stock-work of cassiterite and quartz with topaz (greisen deposit). Due to the high hardness of the greisens, about 50 % of the tin ore of the placer is in aggregates, which must be crushed.

The placers are (except for the Yasny deposit) economic and have the following average parameters: Odinoky stream: extension 5,900 m and average width 640 m. The peat thickness above the placer averages 4.28 m. The average thickness of the productive stratum is 12.62 m. Mokry stream: extension 1950 m and average width 230 m. The peat thickness above the placer averages 4.52 m. The average thickness of the productive stratum is 12.50 m. Slope placer: the area is 907,000 m². The peat thickness above the placer averages 0.12 m. The average thickness of the productive stratum is 5.45 m. The placer’s commercial reserves are 50,900 tons, with an average tin content of 0.83 kg/m³.

Valkumei placer tin deposit

The Valkumei placer tin deposit (Fig. 59) is located at Chaun Bay (Pevek ore placer) on the shore of the East Siberian Sea. It is confined to the offshore flank of the Valkumei tin ore deposit. All the placer bodies are within the tectonic zone separating the arch-dome rise of the Pevek granitoid block and the Chaun neotectonic depression. The placer material sources are mineralized veins and scattered mineralized parts of tin ore/silicate formation confined to the Valkumei block. Outcrops of ore bodies are observed both directly in an abraded cliff and in the area adjacent to the shore belt. The diluvial-alluvial formations deposited at the beach zone also are tin-bearing. The direct tin source for today’s littoral placer is the zone adjacent to the beach in the end of a beach drift N of Cape Valkumei.

Genetically, diluvial, alluvial, littoral and man-made formations are present. The reserves were assessed for a mineable thickness varying from 2 - 50 m (8.8 m on average) with an average tin content of 0.68 kg/m³. The commercial reserves of Category C₁ together with the adjacent manmade placer are 12,500 tons. The, currently, uneconomic reserves (for mining/geological operating conditions > 50 m from the shoreline) are 9,500 t. The inferred resources are estimated to be 35,000 t with an average tin content of 0.7 kg/m³.

Tin and tungsten mining from placer and hard rock ore deposits in the district continued from 1941 till 1992. In 1992 the tin and tungsten mining in the region was stopped due to general economic factors: now the mines and concentration mills are beyond repair, and the mine camps have been abandoned.

Chokurdakh placer tin deposit

The Chokurdakh placer tin deposit is part of the Chokurdakh-Svyatoy Nos placer district located in the offshore Van’kina Bay, south-eastern part of the Laptev Sea, 200 km E of Nizhneyansk settlement. The deposit is confined to the central part
of the Chokurdakh-Lyakhovsky tin-bearing zone and is the first prospected offshore placer tin deposit in Russia. The placer is located at a tectonic zone developed during the Miocene – Holocene on a bed of lacustrine-alluvial loams of Eocene-Oligocene age to an elevation of -60 m, and is represented by tin-bearing, mostly coastal, sediments overlain by Holocene placers, a beach and diluvial sediments. The thickness of the offshore tin-bearing stratum varies from 4 m (littoral) - 58 m (with distance from the shoreline). The placer is 2.4 km long and 520-800 m wide in the centre and 240 m at the flanks. The maximum tin contents (up to 6.9 kg/m³) have been found in the Pliocene-Early Pleistocene layers of the placer’s central part. The average tin content in the deposit is 0.74 kg/m³. The deposit’s reserves are estimated in Category C1 (with all process surveys of sands associated with this estimation level) to be 19,000 tons.

Tirekhtyakh placer tin deposit

The Tirekhtyakh placer tin deposit is also located in Ust-Yana Ulus, 60 km SW of Deputatsky settlement, or 130 km along the winter road via the Mamont deposit. Tirekhtyakh is a unique deposit in Russia, both with respect to reserves and the quality of its sands. The ores and sands of the Tirekhtyakh deposit are free-milling; the concentrates obtained are of a high quality (60 % and higher) and free of harmful impurities, which ensures a high price both in and outside Russia.

As of 01.01.2014 the RF State Register recorded for the Tirekhtyakh tin placer deposit Commercial reserves for open mining, Category B+C1 – 84,682,000 m³ of sands and 68,942 t of tin; for Category C2, 6,638,000 m³ of sands and 5,326 t of tin; non-commercial reserves of Category C1+C2 – 49,044,000 m³ of sands and 14,302 t of tin. Associated component reserves in marketable tin concentrates included the following quantities for Category C1: Commercial: 84,682,000 m³ of sands and 738 t of tungsten trioxide with a grade of 1.34 %, 1.2 t of indium with an average grade of 0.0021 %; the average grade for scandium is 0.223 % and for niobium pentoxide 0.0047 %. Non-commercial resources are 48,336,000 m³ of sands, 219 t of tungsten trioxide and 0.3 t of indium.
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Kolmozero

Polmostundra

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Large diamond deposits are found in kimberlite pipes in the East Siberian minerogenic province, within which several kimberlite fields are distinguished. The main deposits are: Daldyn-Alakit, Lesser Botuoba, and Middle Markha, in which kimberlite bodies of different ages form extensive fields (Fig. 60). The oldest, small dykes and pipes of Precambrian age, are in the southwestern part of the province and contain few diamonds. All primary deposits of economic interest in the province were formed during the period of the Palaeozoic tectonic and magmatic activity; Mesozoic kimberlites have low contents of diamond.

The first kimberlite pipe to be discovered in the USSR (Russia) was the “Zarnitsa” pipe in the East Siberian diamondiferous province (Fig. 60). It was found in bedrock in 1954 as a result of a panning survey focussing on pyrope, led by VSEGEI employees L.A. Popugaeva and N.N. Sarsadskikh, who were the pioneers in this field. In subsequent years, a large number of diamondiferous kimberlite pipes, containing rich diamond deposits were discovered (in several parts of the world) using this method.

"Udachnaya" pipe deposit

The "Udachnaya" deposit is located within the East Siberian minerogenic province in the northern part of the Daldyn-Alakit diamondiferous area in the Daldyn kimberlite field in the junction zone of the Anabar anticline and the Tunguska syncline, to the north of the town of Udachny. The deposit is unique due to its size and average diamond grade. The kimberlite pipe was discovered in 1955, also as
a result of panning prospecting by N. Zdota and V.N. Shchukin of the Amakinsky Expedition and by L.A. Popugaeva and N.N. Sarsadskikh (VSEGEI).

The geology of the area includes a basement of crystalline schists and Archaean granite-gneisses, covered by sedimentary rocks consisting of Ordovician and Silurian terrigenous-carbonate and carbonate rocks, Upper Palaeozoic terrigenous rocks, Upper Permian and Lower Triassic igneous and igneous-sedimentary rocks, and Quaternary sediments. The “Udachnaya” pipe is one of the largest pipes explored in the Republic of Sakha (Yakutia). Since 1971, the deposit has been mined in the “Udachny” open pit. The current surface dimensions are 2.0 × 1.5 km and the deposit has been exploited to a depth of more than 600 m (Fig. 61). The deposit is currently mined from the “Udachny” underground mine. The mine is predicted to reach a maximum production capacity of 4 Mt/a by 2019. The “Udachnaya” pipe intersects the Vendian-Palaeozoic terrigenous-carbonate rocks of the sedimentary cover. It is confined to the intersection of a near-E-W trending fault system with a NW-trending fault (Kharkiv, Zinchuk, Kryuchkov, 1998), and can be traced as a consistent ore body from the surface to a depth of 250 m (Fig. 62). Below this level the pipe splits into two independent ore bodies, Eastern and Western bodies, separated by Upper Cambrian sediments. The distance between the ore bodies increases significantly with depth, from about 100 m at the level of the bottom of the existing open pit (-320 m) to 325 m at -1,080 m elevation.

The Western body is elliptical in shape and elongated in a northwesterly direction. Contacts between the ore body and its host rock are distinct, only rarely becoming so-called “floating” due to host rock brecciation in the proximal excontact zone (Fig. 62). Dip angles in the western part are 80-85°; at the eastern edge, above -789 m elevation, the dip is 60°, and down to -1080 m elevation, the dip steepens to vertical. The Western body contains karst cavities filled with breccias, brine, and gas.

The Western body is, to a depth of 450-530 m, composed of grey and green kimberlite breccias (the first phase of the intrusion). The breccia structure is crystal-lithoclastic, in areas autolithic. The rock is intensely altered by secondary processes. Pseudomorphs of first-generation olivine (xeno- or phenocrysts) amount to 20-25 %; epigenetic material is represented by fragments of sedimentary rocks, crystalline schists and, rarely, ultramafic rocks including garnet (Fig. 4a). Autoliths amount to 26 % of the rock volume; various xenoliths act as autolith cores or centres. The breccia cement consists of carbonate-serpentine aggregate and contains small segregations of ore minerals and numerous pseudomorphs on second-generation olivine.

Kimberlite breccias of the second phase of the intrusion fill the main volume of the deep pipe levels. They are characterized by increased content of pseudomorphs on olivine (up to 30 %), autoliths (up to 25 %), sedimentary rock xenoliths (up to 25 %). The bulk of the rock is replaced by serpentine-carbonate aggregate (Fig. 63). Porphyritic kimberlites with a low content of sedimentary rock xenoliths (3-5 %) were formed at the final stage of pipe development and are present at depths of 294-409 m. They are characterized by the presence of autoliths with increased content of phlogopite in the matrix.

The Eastern body is ellipsoid in plan, elongated in a northeasterly direction and dips at 80-90°. It consists of autolithic kimberlite breccia. Porphyry kimberlite relics occur within the diatreme selvage. Five kimberlite varieties are distinguished in the Eastern body, formed as a
result of four injection phases: 1) Breccia with massive cement texture forming the upper pipe horizons to a depth of 400 m; 2) Autolithic breccia developed in marginal areas of the body below a depth of 350 m; 3) A subvertical stock-shaped breccia body in the central part of the pipe; 4) Breccia in near-contact areas, stretching for almost the entire explored depth of the pipe as a series of small, parallel subvertical injections; 5) A porphyry kimberlite dyke (Kharkiv, Zinchuk, Kryuchkov, 1998).

The content of deep-sourced xenoliths is higher in both bodies of the "Udachnaya" pipe, compared to kimberlites from other fields. In the Western body their average content is 0.1-0.3 %. The most common group (57.1%) of xenoliths are cataclastic garnet serpentinites (apolherzolite variety): Equigranular garnet serpentinites (apodunites, apoharzburgites, apolherzolites) amount to 31.1 %, while garnet-free (spinel apolherzolites) are present in a subordinate amount. Eclogite and pyroxenite are rare, not
exceeding 6%. Kyanitic, coesitic, chromium-pyropic and picroilmenitic varieties of eclogite xenoliths are found. Single grosspydite xenoliths, including diamondiferous ones, alkremites (rocks of spinel - pyrope composition), pyroxenites with primary mica and other rare rocks are found. A large number of deep rocks with diamond-paragenesis minerals are discovered. The only primary minerals preserved in the deep-sourced xenoliths are garnet, picroilmenite, and chromite: the other minerals are replaced by aggregates of serpentine and chlorite.

The composition of the kimberlites, the grain size and the grade of the "Udachnaya" pipe diamonds have been studied in detail, based on data from the mined horizons in the upper part of the deposit. The price of 1 carat diamond of +0.5 mm class from the "Udachnaya" pipe deposit amounted in 2012 to US $ 65.5/carat. The largest diamonds found in the deposit are "Alexander Pushkin" – 320 carats, "The Power of the Soviets" –196.6 carats, "60 years of the Yakut ASSR" – 173.7 carats, "Academician Sakharov" – 172.5 carats.

"Mir" pipe deposit

The "Mir" kimberlite pipe is located within the Lesser Botuoba diamondiferous district of the East Siberian minerogenic province (Fig. 60). It was discovered on June 13, 1955 by geologists of the Amakinsky Expedition - Yu.I. Khabardin, E.N. Elagina, and V.P. Avdeenko. The rocks that host the "Mir" pipe are sub-horizontally bedded and of Cambrian age. The "Mir" pipe is confined to the approximately N-S-trending zone of the Parallel fault of the Vilyui-Markha group of regional tectonic faults. The pipe-shaped kimberlite is steeply plunging, and has a conical shape to a depth of 300 m (elevation +30 m), and becomes cylindrical in the depth interval of 300-900 m (elevation range of +30 m/-600 m). The body becomes markedly narrower in the depth range of 900-1,000 m and evolves into a subvertical kimberlite feeder dyke, about 300 m long and 25-30 m thick. The "Mir" pipe is composed of kimberlitic rocks formed as a result of a three-phase kimberlite magma intrusion.

Rocks of the different phases do not show significant variation in composition, physical and mechanical properties, or diamond grade. They include kimberlite breccias containing variable amounts of carbonate host rock and dolerite fragments. Their distribution in the pipe is uneven (Fig. 64).

Upper mantle xenoliths are widespread in the "Mir" pipe. Pyrope lherzolite prevails (Fig. 65); diamond-bearing ultramafite and eclogite are found and also relatively low-pressure pyroxenite, spinel and spinel-free ultramafite, mica pyroxenite and cataclastic varieties of these. Their distribution pattern in the pipe has not been determined. Inclusions of other minerals in the diamonds belong to ultramafic (99.45 %) and eclogitic parageneses.

Diamond recovery in the field began in 1957 by open-pit mining and continued for 44 years. The...
open pit in the "Mir" pipe field is 525 m deep and 1.2 km in diameter. Open-pit mining of the diamondiferous kimberlite ceased in June 2001. According to exploration results, the depth of diamondiferous kimberlites in the field is more than a kilometre. "Alrosa" started underground diamond recovery in 2009. The diamond content of the "Mir" pipe is high, averaging above 3 ct/t. The diamonds are noted for their fairly high quality; the average price for diamonds from the deposit amounted in 2012 to US $ 94.96/ct. The diamonds include octahedra (61 %), rhombic dodecahedra (10 %), combinational forms (30 %), and cubes. Colourless stones are most common but brownish, bluish-green, smoky gray, and purple varieties also occur.

Studies of the diamond content in the pipe area and to a depth of 1,000 m have not shown any regular variations in their distribution in kimberlite ore. The largest diamond, "XXIV Congress of the CPSU" recovered at the "Mir" mine in 1980 weighed 342.5 carats. The reserves of the "Mir" pipe deposit in A+B+C1 grade are 139,558.9 Kct, in C2 grade, 3,338.5 Kct.

"Internationalnaya" pipe deposit

The "Internationalnaya" kimberlite pipe lies within the East Siberian minerogenic province in the Lesser Botuoba diamondiferous district, 16 km SW of the "Mir" pipe (Fig. 60). The pipe was discovered in 1969 by A.G. Ivanov, V.F. Romanov, A.A. Gorbunov, I.N. Ivanov, V.N. Shchukin, A.A. Vasiliev, V.S. Mokeev, I.A. Pogudin, and M.I. Popov.

The kimberlite pipe is accompanied by a system of dykes and is overlain by a thin (up to 9.2 m thick) stratum of Lower Mesozoic sediments. The host rocks are Cambrian and Lower Ordovician sediments forming a monocline. The pipe is 0.4-0.5 km W of the axis of the ore-controlling Kyuellyakh fault. The pipe dimensions do not change significantly to a depth of 1,000 m. It is a funnel-shaped body to a depth of 125 m which evolves into a near-cylindrical body plunging steeply southeastwards. At the surface the pipe has an irregularly oval cross-section, elongated in northwesterly direction, with dimensions of 152 × 112 m. At the -560 m level, the dimensions of the pipe are 104 × 70 m, at the -690 m elevation, 99 × 68 m, and at the -820 m elevation, 92 × 62 m.

The upper horizons of the pipe comprise kimberlite breccias cutting massive porphyritic kimberlites. The kimberlite breccias contain, to a depth of 370 m, considerable amounts of wall rock (sediments). In general, the ore body is represented by autolithic kimberlite breccia (80 %) composed of round, oval kimberlite segregations of early-generation (lapilli) and porphyritic kimberlite varieties. The kimberlite rocks of the pipe are characterized by low contents of heavy minerals. Pyrope prevails over picro-ilmenite;
unaltered olivine, chrome diopside, and zircon are rare. In its content and the composition of deep-sourced minerals, the kimberlites of the pipe differ from the vast majority of the bodies in Yakutia.

The autolithic and porphyritic kimberlites are generally similar in terms of diamond grade, but the average grade is slightly higher in the porphyritic kimberlites. The average diamond grade decreases with depth. The diamonds that dominate in this deposit have an octahedral habit, with a low content of twinning, splices, and naturally stained crystals. The crystals have high clarity and the number of coloured crystals and stones with solid inclusions is low. As a result, the diamonds in the deposit are of high quality and value. In combination with the high diamond grade (average ca. 8 ct/t of ore) this makes the "Internationalnaya" pipe deposit globally unique in terms of diamond material value. The average price for diamonds from the deposit in 2012 amounted to US $ 145.72/ct. Balance deposit reserves in A+B+C1 grade are 37,020.9 Kct, in C2 grade, 12,303.7 Kct.

"Botuobinskaya" and "Nyurbinskaya" pipe deposits

The "Botuobinskaya" and "Nyurbinskaya" pipe deposits and related placers of the same names occur within the Middle Markha diamondiferous district and are confined to the Nakyn kimberlite field in the southern part of the East Siberian province (Fig. 60). The geology of the area is determined by its location in the central part of the Siberian Platform at the junction of the southeastern slope of the Anabar anticline and the Vilyui syncline. The "Botuobinskaya" kimberlite pipe was discovered in 1994. It is 3.3 km southwest of the "Nyurbinskaya" pipe. Open pit mining started in 2012. Ore and sand extraction should be launched in 2015. The "Nyurbinskaya" pipe was discovered in 1996. The "Nyurbinsky" open pit was commissioned in 2000 and as of 01.07.2013 its depth was 255 m.

The stratigraphic succession of the host rocks includes Upper Cambrian and Lower Ordovician strata, into which the kimberlites were emplaced. Some 60-70 m of Triassic and Lower and Middle Jurassic strata overlay the kimberlites. The kimberlites are confined to the intersection of the Vilyui-Markha and Middle Markha major fault zones. Palaeo-depressions and sink holes, mostly filled with unconsolidated sediments of Dyakhtar (Triassic-Lower Jurassic), age are common in the area of the primary deposits and the diamond content in the basal layers of the Ukugutian Formation of the Lower Jurassic is such that they are of commercial interest as placer diamond deposits.

The "Nyurbinskaya" pipe has a northeasterly strike and is confined to the axial line of the Dyakhtar fault. In plan view it has a rounded-ellipsoid shape. The dimensions of the upper part of the pipe (+190 m elevation) are 358 × 177 m, with an area of 41,700 m². The pipe width decreases significantly with depth and at -55 m elevation its horizontal intersection is 24,220 m², and at -200 m elevation, only 1200 - 1800 m². At a depth of 280-320 m below the surface, the pipe is divided into two ore bodies separated by a dyke-shaped mafic intrusion. The contacts of the pipe with its wall rock have a general dip of ca. 80° towards the central part of the pipe. Kimberlite injections up to 0.5 - 3 m thick occur along the contact of the pipe and into the wall rock. Residual weathering crust represented by altered breccias and clayey greenish-gray formations is observed in the upper horizons of the pipe.

The "Nyurbinskaya" pipe is the largest pipe within the Nakyn kimberlite field. It hosts three types of kimberlite: autolithic kimberlite breccias, kimberlite breccias comprising the main pipe volume, and porphyry kimberlites that have limited distribution in the deeper levels. Fine- to medium-grained clastic breccias in the central part of the ore body and carbonate kimberlite breccia in the near-contact zone of the northeastern and southwestern flanks of the pipe are distinguished among autolithic kimberlites. Autolithic breccias of the central part of the pipe consist of fine- to medium-grained porphyry rocks with an autolithic structure in the kimberlite cement. They are characterized by the presence of small amounts of wall rock debris, giving it the breccia texture. The content of xenoliths of metamorphic rock reaches 10 %. Xenoliths of mantle rocks are extremely rare (0.1 %) and are represented by fragments of garnet serpentinite and glimmerite. The heavy fraction of the kimberlite pipe, is dominated...
by pyrope and chrome-spinel, while picro-ilmenite, olivine, and clinopyroxene are rarer. These minerals are generally fine-grained, with a maximum dimension of 5 mm. The pipe has a high diamond grade; the average content in the reported reserves of the field exceeds 4 c/t. The diamonds predominantly show octahedral and transitional octahedral-rhombohedral dodecahedral crystals (up to 87 % of the total amount). 5-10 % of the diamonds are yellow-green crystals, rarely gray. The composition of the kimberlites, their grain size and diamond grade have been studied in full, based on data from mining of the upper horizon. According to the price list of the Ministry of Finance of the Russian Federation, the average price for 1 carat diamond in +3 nominal sieve class from the "Nyurbinskaya" pipe mine in 2012 was US $ 56.2/ct.

The "Botuobinskaya" pipe is a complex, paired kimberlite body. The field hosts two kimberlite phases: porphyritic, composing the dyke part of the pipe, and explosive autolithic kimberlite breccias of the second phase developed mainly in the vertical channel. Crater facies rocks are preserved in the upper part of the pipe (Fig. 66).

The diamond grade of the "Botuobinskaya" pipe is very high, averaging about 5 ct/t. The pipe contains many gem and near-gem quality crystals with a high transparency and low proportion of coloured stones. Fragmented crystals prevail but the quality of the stones is still high. The average price for diamonds from the deposit is US $ 72/ct.

Lomonosov diamond field

The Lomonosov diamond field lies within the East European minerogenic province, in the Arkhangelsk diamond subprovince, Zimny Bereg diamondiferous area. The discovery in the early 1980s of a new diamondiferous kimberlite subprovince in the north of the European part of Russia (100 km NE of Arkhangelsk) was the result of a systematic study of the geological structure of the region, started by Arkhangelsk geologists in the early 1960s. In 1978, an airborne magnetic survey was conducted, promising anomalies were distinguished and test drilling intersected a large number of diamond-bearing pipes, including commercial ones, forming individual fields (Fig. 67). The pioneers in exploration of the Lomonosov diamond field, which comprises six high diamondiferous kimberlite pipes, were A.F. Stankovsky, E.M. Verichev, K.N. Sobolev, A.V. Efimov, V.P. Grib, V.A. Medvedev, V.A. Larchenko, Yu.G. Konstantinov, L.P. Dobeyko, S.P. Aleksandrov, V.S. Fortygin, G.Z. Grinevitsky, V.A. Lyutikov, R.S. Kontorovich, and G.M. Levin.

The kimberlite magmatism of the Arkhangelsk subprovince is associated with the Middle Palaeozoic tectonomagmatic activation of the East European province. The reserves of the district, which hosts the kimberlite pipes "Arkhangelskaya", "Karpinsky-1", "Karpinsky-2", "Lomonosov", and "Pionerskaya", were estimated in 1987.

Structurally, the field is within the Tovsky block of the crystalline basement (Arkhangelsk diamond province, 1999) and is characterized by a linear-chain arrangement of pipes in the area of a N-S trending deep fault. The total length of the pipe chain is 14 km and the distance between the individual pipes varies from 100 m to 2.5 km. The pipes give weak magnetic anomalies with
intensities from 15 to 50 nT. They intrude into Upper Vendian terrigenous sedimentary rocks. Isometric subsidence troughs exceeding the pipe area by 6-8 times are observed around the pipes in the top Vendian strata (Erichenk et al. 1997). The average thickness of the overburden layer varies from 28.3 m ("Arkhangelskaya" pipe) to 54.5 m ("Lomonosov" pipe).

Each pipe consists of a feeder channel, which, in the case of the "Arkhangelskaya", "Karpinsky-1", and "Pionerskaya" pipes is overlapped by crater facies. The "Arkhangelskaya", "Karpinsky-1", and "Lomonosov" pipes are almost circular in plan view, while the "Karpinsky-2" and "Pionerskaya" pipes are elliptical (Fig. 68). The thicknesses of the crater facies range from 72 m ("Karpinsky" pipe) to 123 m ("Arkhangelskaya" pipe) (Verichev, Garanin, Grib et al., 1991).

The age of the kimberlite, determined both from geological data and from the age of carbonized wood from xenoliths in the kimberlite breccia, corresponds to the Middle Palaeozoic. The diatremes are composed of rocks of crater, vent, and hypabyssal facies. Rocks of the crater facies are represented by tuff, tuffite, tuff sandstone and tuff siltstone, and sedimentary breccia in the near-contact parts. The rocks are composed of saponite pseudomorphs on olivine and autolithic lapilli (10-20 %), but mostly of wall rock and individual quartz grain fragments.

Vent facies formations are represented by xenon- and tuff breccias. Rock textures are brecciated; lapilli proportions vary from 10 to 60 % and are unevenly distributed in the rock. Wall rock fragments are up to 20 cm in size (Fig. 69). Lapilli are represented by crystal-clasts of altered olivine and spherical autoliths and xenoliths of the enclosing rocks. Porphyritic impregnations are represented by megacrysts and phenocrysts of altered olivine and rare phlogopite. The matrix consists of fine aggregate of the same components, with an abundance of xenogenic quartz grains cemented by cryptocrystalline aggregates, which include saponite, illite, carbonate, chlorite and iron hydroxides (Lukyanov, Lobkova, Mikhailov et al., 1994).

Mantle rock xenoliths are rare and intensely altered (Sobolev, Pokhilenko, Grib et al., 1997). Autolithic breccias of the vent facies consists of a variable content of lapilli with megacrysts.
of altered olivine and spherical autoliths. Mantle xenoliths and rocks from the basement and sedimentary cover are rare. All of these components vary greatly in abundance in different parts of the pipes and in the various pipes. In some cases (e.g. at a depth of 800 m in the Pionerskaya pipe) autolithic breccias grade into massive porphyry kimberlite. The pipe's roots consist of rocks of the hypabyssal facies composition, represented by porphyritic and aphanitic kimberlites. Typical porphyry kimberlite consists of elliptical megacrysts (size up to 10 mm, over 30 %) and olivine impregnations (grain-size <1.5 mm, up to 30 %) embedded in a groundmass composed of olivine microlites, phlogopite laths, fine-grained oxide, calcite microlites, and a matrix of crypto-crystalline serpentine-chlorite aggregate.

The diamond crystals are usually gray or with a gray flash (42 %). The proportion of colourless diamonds is 39 %. Coloured (black, yellow, green-gray, brown) crystals and those with different shades account for 19 %. More than 50 % of the crystals are transparent. The proportion of opaque diamonds is 15 %, that of splices 21.5 %, of crystal fragments 11.8 % and twinned crystals 10 %. Forty percent of the crystals are fractured. The proportion of isometric crystals is 35 %, while 25 % are deformed.

Lomonosov GOK, the production enterprise of the Open Joint Stock Company "Severalmaz", is currently developing the "Arkhangelskaya" and "Karpinsky-1" pipes. Open pit mining was initiated on the "Arkhangelskaya" pipe in 2005. The surface dimensions of the quarry are 1.16
km × 1.12 km, and its depth is currently 110 m. Stripping operations began at the "Karpinsky-1" pipe field in 2010. Currently, the quarry depth is 90 m; its surface dimensions are 830 m × 500 m. The "Karpinsky-1" quarry is scheduled for full production capacity (2 Mt/a) in 2015.

V. Grib diamond field

The V. Grib diamond field consists of a single kimberlite pipe located within the Arkhangelsk diamond subprovince. It was discovered in 1996. It is located in the centre of the Zimny Bereg diamondiferous area, 30 km from the Lomonosov diamond field (Fig. 67). Diamondiferous kimberlites were uncovered by drilling in the area of a magnetic anomaly with an intensity of 15 nT. 150 diamond microcrystals were found in the kimberlite core after thermochemical digestion. The pioneers in this field are A.N. Buyun, E.M. Verichev, N.N. Golovin, A.A. Zaostrovtsev, V.F. Kurushin, and V.I. Sotnikov.

The kimberlite pipe intrudes weakly lithified Vendian terrigenous sediments; it is covered by Early Carboniferous rocks and Quaternary deposits with a total thickness up to 70 m. The pipe show features typical for weakly eroded pipes, with a clearly manifested crater and vent components. Its diameter is 1.6 km (Fig. 70).

The petrographic composition of the crater facies is defined by the ratios of magma, silty-sandy material and xenoliths of the enclosing terrigenous rocks. The magmatic material in it, unlike the vent, is significantly altered. The cement is iron-clayey, carbonate, or saponite. The vent part of the pipe is composed of rocks of two injection phases – tuff- and xenotuff breccias gravitating towards the southern part of the pipe and massive kimberlites of a later injection phase. The kimberlites are characterised by a lithoclastic, psammitic to fine-psephitic structure and massive texture. The crystalloclasts are of olivine or olivine pseudomorphs. Olivine of the first generation (38-46 %) is generally 1-4 mm in size (up to 2.5 cm) and has a rounded shape. Olivine of the second generation (2 %) is represented by idiomorphic crystals 0.1-0.8 mm in size.

Cystalloclasts of pyrope, pyrope-almandine, ilmenite, phlogopite, and clinopyroxene do not

<table>
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<th>Diamond Grade +3 nominal sieve class (ct/t)</th>
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</table>


Figure 70: Schematic section of the Grib kimberlite pipe (Arkhangelsk diamond province, 2000). 1: Quaternary deposits (loam, sand); 2: Carboniferous limestone, dolomite; 3: Urzugian Formation sandstone; 4-7: Crater facies rocks: 4: Clayey sand; 5: Tuff sandstone; 6: Tuff and tuffite; 7: Host-rock breccias, conglomerates with tuff and tuffite inclusions; 8-9: Vent facies rocks: 8: Tuff-xenotuff breccias; 9: Kimberlite; 10: Vendian host rocks; 11: Drillholes.
exceed a total of 4%. Lithoclasts (10%) are represented by nucleated, zoned autoliths of oval shape and 1-30 mm in size. The autolith structure is porphyritic: second generation olivine impregnations make up 25-30% of the autolith volume; phlogopite, not exceeding 5%, is represented by fused laths. Oxides in the autolith matrix are perovskite, titanomagnetite, chrome spinel, and ilmenite. Xenogenic material (3.1%) includes fragments of mudstone, siltstone, and plutonic rocks. The dimensions of the latter sometimes reach 20 cm. Granulite and eclogite-like rocks are most common, while pyrope and ilmenite, as well as various altered phlogopite rocks are rare. Mantle rocks are dominated by pyrope-bearing dunite and lherzolite; less common rocks are olivinite, pyrope-bearing clinopyroxenites and websterites. The rock cement (30-40%) is composed of serpentine with a small carbonate admixture. The kimberlites in the V. Grib pipe differ from the Zolotitsa field kimberlites in their high content of kimberlite indicator minerals and in the predominance among them of picroilmenite, garnet, and clinopyroxene and lesser amounts of chrome spinel.

Diamonds from the V. Grib pipe are characterised by a large proportion of high clarity stones (80% of the total). The diamond crystals are predominantly octahedral (37%); dodecahedral crystals (26%), and crystals of transition type O - D (26-20%). Assessment and approval of the reserves was carried out in 2005 and additional exploration was completed in 2010. Currently the diamonds are being mined by open pit methods; the depth of the pipe revealed by the pit is 136 m.

<table>
<thead>
<tr>
<th>Object</th>
<th>Component</th>
<th>Units of ore measure</th>
<th>Units of comp. measure</th>
<th>Ave. grade</th>
<th>Units of ave. grade measure</th>
<th>ABC1 (ore)</th>
<th>ABC1 (com)</th>
<th>C2 (ore)</th>
<th>C2 (com)</th>
<th>Off-balance (ore)</th>
<th>Off-balance (com)</th>
</tr>
</thead>
<tbody>
<tr>
<td>V. Grib pipe</td>
<td>Diamond</td>
<td>Kt</td>
<td>Kct</td>
<td>1.245</td>
<td>ct/t</td>
<td>56,677</td>
<td>70,581</td>
<td>19,395</td>
<td>14,614</td>
<td>33,379</td>
<td>12,473</td>
</tr>
</tbody>
</table>

Table 7: Reserves at V Grib pipe as of 01.01.2014.

**PLACER DIAMOND DEPOSITS**

"Solur-Vostochnaya" placer diamond deposits

The "Solur-Vostochnaya" placer deposit is located in the East Siberian minerogenic province. It consists of two ancient, spatially separate, buried deposits "Solur" and "Vostochnaya" located in the area between the Irelyakh and Chuonalyr rivers, 25 km NW of the town of Mirny. The deposit area is 15.18 km². The deposits were discovered during prospecting work undertaken by the Botuoba Expedition in 1979 and 1989, respectively. In 1981-2005, exploration and reserve estimation work was completed on the deposits.

The sedimentary formations which host the diamond-bearing primary deposits in the area are terrigenous-carbonate rocks of the Vendian-Upper Cambrian periods. The overlying terrigenous deposits of the Upper Palaeozoic, Mesozoic, and Cenozoic host the diamond placers. Rocks of the Middle Palaeozoic kimberlite formation, belonging to the Mirny field, are situated to the southeast (12 km) and east (20 km) of the "Solur - Vostochnaya" placer deposit. They form a series of kimberlite pipes and bodies, including the "Internationalnaya" and "Mir" pipes (described above). The pipes are characterized by a high diamond grades and are the main sources for the Upper Palaeozoic and Mesozoic alluvial deposits within the area.

The "Solur-Vostochnaya" field consists of two spatially separate but contiguous buried strata; the Middle Carboniferous "Vostochnaya" strata...
and the Mesozoic "Solur" strata. "Vostochnaya" strata are localized within the Ottursky valley-like palaeodepression with a northwesterly strike, 5 km in length and 1-2 km wide. The "Solur" Early Jurassic strata are controlled by the smaller Solur palaeodepression with a southeasterly strike.

The "Vostochnaya" deposit, with its alluvial genesis, is located within the coarse clastic basal horizon of the Middle Carboniferous Lapchanskian Formation. The economic circumference is 4.6 km. The productive mantle-like horizon varies in thickness, from 0.1-1.9 m, averaging 0.68 m. The lithological composition of the layer includes conglomerate, pebbles, sand and sandstone with admixed gravel and pebble material, clayey sand, sandy siltstone with gravel and clayey siltstone. The thickness of sediments overlying the productive layer in the "Vostochnaya" deposit varies from 12-58 m, averaging 47.7 m.

The "Solur" deposit, with its alluvial genesis is confined to the basal level of the Lower Jurassic Yulegir Formation. Upper Cambrian terrigenous-carbonate rocks commonly form the placer basement. The thickness of the productive layer varies from 0.5 to 5.1 m, with an average of 2.35 m. The productive layer is mainly confined to the basal alluvial pebble-rich bed with interbeds and lenses of gravelite, clayey sand and sandstone, siltstone and silty clay. The rocks of the productive layer are poorly lithified. The thickness of the sediments overlying the "Solur" deposit is 5-54 m, averaging 41.7 m.

In the "Vostochnaya" placer the content of diamonds in individual samples varies from 0.0 - 99.29 ct/m³, with an average of 3.01 ct/m³ for the deposit as a whole. The diamond distribution is irregular; enriched areas are found mainly associated with the conglomerates in the central part of the deposit. The diamond content of the "Solur" placer is considerably lower, with samples varying from 0.0-10.12 ct/m³, with an average of 1.0 ct/m³. Diamonds from the two areas are similar. The dimensions of the diamonds in both placer deposits vary widely, but diamonds of -4 +2 mm and -2 +1 mm classes predominate: their total amount, in general over the field, is 94.7 %. -4 +2 mm class is dominant by weight; its content in general over the placer field is 66.4 %. The average weight of 1 crystal in +1 mm class over the Vostochnaya deposit is 17.1 mg, in the "Solur" deposit as a whole, 20.3 mg (Fig. 71).

The habit and morphological character of the diamonds from the "Solur-Vostochnaya" deposit are identical to those of the "Mir" pipe and the "Vodorazdelnye Galechniki" placer. 34 % of the "Solur-Vostochnaya" field diamonds have

Figure 71. "Solur" placer diamonds (Grakhanov, Shatalov, Shtyrov, 2007).
gem quality. The price of 1 carat diamond of +1 mm class from the "Solur-Vostochnaya" placer amounted to US $ 96.18/carat. The Balance reserves in A+B+C grade are 15,903.2 Kct and in C₂ grade 864.7 Kct.

**Ebelyakh River and Gusiny Stream placer diamond deposits**

The Ebelyakh diamond area, covering the Gusiny Stream placer deposit is located in the Anabar-Olenek diamondiferous region, in the East Siberian minerogenic province, in the northeastern margin of the Siberian Platform, at the junction of the Anabar anticline and the Lena-Anabar deflection. The area was discovered in 1965. The pioneers in finding these placer deposits were the geologists of the Amakinsky Expedition, Yu.P. Belik, S.A. Grakhanov, L.M. Zaretsky, V.M. Kunitsky, Y.A. Lomakin, V.M. Podchasov, and M.A. Chumak.

The area is represented by terrigenous-carbonate and volcanic rocks of the Middle Cambrian, Carboniferous, Permian, Triassic, Jurassic and Cretaceous, as well as polygenetic unconsolidated sediments of Cenozoic age. Diamonds are found within terrigenous rocks from the Permian, Jurassic and Lower Cretaceous periods in the weathering crusts and in all unconsolidated formations of Neogene to Quaternary age. (Grakhanov, Shatalov V.I., Shtyrov V.A., et al., 2007).

The Ebelyakh River placer covers the river valley section stretching for 83 km from its mouth. The valley cuts through Middle Cambrian carbonate rocks, which are represented by dolomite of the Anabar Formation and limestone of the Dzhakhtar series. The depth of the valley with respect to the watershed averages 100 m, varying from 110 m near the mouth to 80-90 m in the upper reaches of the river. The valley slopes are terraced in the lower and upper reaches of the river; but are virtually absent in the central section.

The diamond field is a large alluvial placer; its economic circumference includes the sediments of the river bed, lower and upper plains, four flood-plain terraces, and redeposited weathering crusts. The commercial placer has a length of 82 km. Its width averages 78.6 m, varying from 50 to 345 m. According to the conditions, occurrence, size and degree of continuity of the stratum, and the distribution of the commercially valuable sections, the deposit can be divided into two areas: the valley and the terraces. The valley areas of the deposit have relatively continuous widths and thicknesses with relatively flat-lying sediments with a shallow dip. The distribution of diamonds is highly variable. Terrace placer areas have been formed from fragments of previously large, partially eroded placer deposits. These are characterized by relatively continuous widths and thicknesses, within which relatively broad zones can be distinguished as having low diamond grades. The main lithologies forming the river bed and upper and lower plains are pebble-gravel-sandy sediments, silts with pebbles and rock debris. The diamond-bearing layer in the terraces consists of pebble-gravel-sandstone sediments, boulder-pebble-gravel formations and alluvium underlain with redeposited weathering crusts. The bedrock of the alluvial complex in the Ebelyakh River valley is composed of either carbonate rocks from the Middle Cambrian or of weathering crusts.

Based on the occurrence of the diamond-bearing horizon, size, discontinuity, granulometric composition of diamond-bearing layers, regularity of diamond distribution, the following three sections of the Ebelyakh River placer deposit can be distinguished: "Priustyeyuv" (Confluence area), "Nizhny" (Lower area) and "Verkhny" (Upper area). The average diamond grade within the Ebelyakh River placer deposit varies along the exploration lines from 0.28 to 7.82 ct/m³, while along the selected estimated blocks the variation is from 0.22 to 5.91 ct/m³. The average value across the deposit is 1.34 ct/m³.

**Gusiny Stream**

The Gusiny Stream placer deposit is genetically related to the alluvial sediments. The economically mineable portion of the placer extends for 8.7 km. Its width along the trend of the deposit varies from 42 - 262 m, with an average of 128 m. The thickness of the diamond-bearing layer is quite uniform, on average 2.20 m. It is covered by a 2.2 m thick peat layer. Two sites are distinguished within the alluvial deposit, the "Verkhny" (Upper) site and the "Nizhny" (Lower) site. These are characterized...
by different diamond contents, thickness of the diamond-bearing layer, width of the commercial circumference and the stripping ratio. The average diamond grade within the Gusiny Stream placer deposit varies along the exploration lines from 0.38 to 2.40 ct/m³: block grades vary from 0.74 to 1.87 ct/m³ and the average value for the whole deposit is 1.364 ct/m³. About 170,000 diamond crystals with a total weight of 17.74 Kct were extracted in the process of geological and exploration work within the Ebelyakh River, Gusiny Stream, and Yraas-Yuryakh Stream (another tributary of the Ebelyakh River) area. Studies of the crystals revealed that the size of diamonds from the Ebelyakh River placer deposit varies from fine crystals to large stones in excess of 20 ct. The “Priustyeyvo” site (Confluence area) is characterized by larger diamonds. Generally the size of the diamonds in the deposit gradually decreases upstream relative to the Ebelyakh River. Study of the typomorphic properties of the diamonds within the Ebelyakh River placer deposit have shown that they are different from the diamonds of the Daldyn-Alakit and Lesser Botuoba kimberlite areas. Industrial-quality diamonds are predominant. The content of gemstone diamonds (category I) comprises only 6.55 % of the total. The deposit is characterized by low-integrity diamonds; their average value is US $ 49.88/ct.

<table>
<thead>
<tr>
<th>Grade</th>
<th>Sands (k m³)</th>
<th>Grade +2 mm (ct/m³)</th>
<th>Contained Diamonds +2 mm (Kct)</th>
<th>Off-balance reserves</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ebelyakh River placer</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>2,436</td>
<td>1.55</td>
<td>3,784</td>
<td>-</td>
</tr>
<tr>
<td>C₁</td>
<td>13,396</td>
<td>1.46</td>
<td>19,494</td>
<td>585</td>
</tr>
<tr>
<td>B + C₁</td>
<td>15,832</td>
<td>1.47</td>
<td>23,278</td>
<td>585</td>
</tr>
<tr>
<td>C₂</td>
<td>3,800</td>
<td>0.76</td>
<td>2,885</td>
<td>1,416</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.34</td>
</tr>
<tr>
<td>Gusiny Stream placer</td>
<td></td>
<td></td>
<td></td>
<td>485.1</td>
</tr>
<tr>
<td>B</td>
<td>804</td>
<td>1.35</td>
<td>1,088</td>
<td>-</td>
</tr>
<tr>
<td>C₁</td>
<td>1,743</td>
<td>1.37</td>
<td>2,383</td>
<td>90</td>
</tr>
<tr>
<td>B + C₁</td>
<td>2,547</td>
<td>1.36</td>
<td>3,471</td>
<td>90</td>
</tr>
<tr>
<td>C₂</td>
<td>1</td>
<td>4.53</td>
<td>2.9</td>
<td>86</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td>0.45</td>
</tr>
</tbody>
</table>

Table 8: The balance reserves of the Ebelyakh River and Gusiny Stream deposits as of 1.01.2013 (“Micon International Co Limited”. 2013).

**IMPACT DIAMOND DEPOSITS**

**Impact diamond deposits of the Popigai area**

Primary and placer impact diamond deposits are found in the Popigai diamondiferous area, within the East Siberian minerogenic province, on the northeastern margin of the Anabar Shield of the Siberian Platform. The source of the diamonds is rocks within the large, ancient Popigai meteorite crater (Fig. 72)

In 1971-1987 and later, a comprehensive study of the Popigai structure was undertaken by VSEGEI experts, geologists of the Kotui Party and Polar Exploration Expedition PGO "Yakutskgeologiya". The discoverer of the
The Popigai impact structure (diameter 100 km) resulted from an asteroid impact approximately 35.7 million years ago. Rocks of the basement (various gneisses and schists) and the cover of Late Proterozoic – Mesozoic sediments with total thickness of about 1 km, including graphite-bearing horizons, became the target for the asteroid. The internal astrobleme structure is characterized by the presence of a central depression, a circular elevation of the crystalline basement, and a ringshaped trench surrounded by a zone of deformed rocks. The depression and trench are filled with different impact breccias and impactites. Impact breccias and impactites are also developed in small areas outside the crater.

Impactites were formed by impact metamorphism and gneiss melting, as well as by intrusion of the melt. Graphite underwent polymorphic transition to diamond due to quasi-hydrostatic compression above 35 GPa. Diamondiferous impactites are represented by two varieties - tagamites and suevites. Impactites are exposed on the surface within an area of approximately 1,140 km²; the total area of their development is approximately 3,500 km²; diamonds are ubiquitous in this area. Diamond-bearing impactites occur as thick (up to 600 m), extensive (up to 10-15 km) sub-horizontal and lenticular bodies; smaller irregular bodies tens of metres thick are also present.

Unconsolidated Pliocene - Quaternary sediments are widespread, serving as intermediate accumulators for placer deposits. Primary
deposits of the Popigai structure are represented by the Udarnoe and Skalnoe fields. The Udarnoe deposit is located in the NW sector of the Popigai structure; covering an area of 8.3 km². Tagamites form tabular bodies from a few tens to >100 m thick and rarely also small lenticular bodies of irregular shape in allogenic monomictic breccia of crystalline rocks. Massive tagamites are widespread. Ataxite (iron meteorite) and porous impact melt are less frequent. Suevite forms a single thick tabular body covered by tagamites: its average thickness is 63 m (up to 264 m), increasing from south to north.

The proven area of productive impactite strata is 7.66 km², its thickness varying from 368.9 to 97.7 m, increasing from the southeast to the northwest. The strata comprise three ore horizons differing in material composition and diamond content: upper tagamite, suevite, and lower tagamite. The vast majority of the tagamite horizon is composed of massive rocks 110.9 m thick; suevites occupy only an insignificant volume (4%). The diamond content in the productive series is from 2.20 - 21.64 ct/t in individual drill holes and, relative to reserve estimation blocks, from 2.63 - 13.64 ct/t. The average content of the deposit in reserve estimates is 7.6 ct/t, of which in tagamites - 9.5, and in suevites - 6.6 ct/t. The cover of eluvial-deluvial deposits with an average thickness of 3.8 m is rich in diamonds - 8.43 ct/t. 90 % of the diamonds are light-coloured varieties (Fig. 73).

The Skalnoe deposit in the southwestern sector of the crater has dimensions of 13×6.5 km and covers an area of about 85 km². It covers the SW slope of a circular high within a ring trench. Impact metamorphosed and cataclastic gneisses of authigenic breccia traced by drillholes to a depth of 900 m are exposed in the axial part of the high. In the SW part of the deposit gneisses are covered by thick lenses of different allogenic breccias cemented by suevites and tagamites: their thickness exceeds 1 km. A flat-lying tabular tagamite body, associated with the highest diamond contents, is traced in the northeastern part of the deposit. The diamond content in the field varies laterally and with depth. Tagamites contain an average of 17.3, and suevites 11.5 ct/t. Marginal (near-roof) parts of tagamite beds are usually enriched in diamonds, the content reaching 24.5 ct/t; rich ore is available for open-pit mining. Light-coloured diamonds amount to approximately 85 %.

**Impact diamond placers**

Impact diamonds in unconsolidated Pliocene-Quaternary sediments are identified as different genetic types in the entire area of the Popigai impact structure and in close proximity to it. The advantage of placers over primary deposits is their higher content of coarse diamond fractions, their greater hardness, as well as cheaper technology for extraction. The diamond content of alluvial deposits in the Popigai basin has been determined for the alluvium in all the river systems. The highest concentrations of impact diamonds in eluvial-deluvial sediments (in some samples up to 166 c/m³) are found in the Skalnoe and Udarnoe deposits. The diamond reserves in the Udarnoe deposit are estimated in B+C1+C2 grades and amount in total to more than 12 Gt (Giga carat) with a ratio of grades, respectively, of 1:2:3. In the Skalnoe deposit area, commercial high-grade reserves amount to approximately 150 Gt. Inferred resources in P1 grade for the Syuryunge, Vstreechny and Tongulakh sites, etc. amount to 50 Gt.

Proven reserves and inferred resources of impact diamonds in bedrock in the total area of approximately 120 km² in Popigai area amount to 212 Gt. The total number of diamonds in an impactite layer 50 m thick in the rest of the area (about 1,020 km²), in addition to the deposit areas and prospects, are assessed to be about 150 Gt. Reserves and resources of placer impact diamonds are very significant: The diamond reserves of the Popigai astrobleme exceed the diamond reserves of all other diamondiferous provinces in the world.

![Figure 73. Impact diamonds collected from the Popigai impact structure (Ohfuji et al. 2015).](image)
REFERENCES


Masaitsis V.L., Kirichenko V.T., Mashchak M.S. et al., 2013: Primary deposits and placers of impact diamonds of Popigai area (Northern Siberia). Regional geology and metallogeny 54, 89-97.


Deposits

- Diamonds
- Hydrothermal fields
- Energy metals: U, Th
- Precious metals: Ag, Au, Pd, Pt, Rh
- Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Base metals: Al, Co, Cu, Ni, Pb, Zn
- Ferruginous: Cr, Fe, Mn, Ti, V

Size and activity

- Active mine
- Important deposit
- Very large with active mine
- Very large
- Large with active mine
- Large
- Potentially large with active mine
- Potentially large

Legend:
- Red diamond: Diamonds
- Blue triangle: Hydrothermal fields
- Green circle: Energy metals: U, Th
- Yellow circle: Precious metals: Ag, Au, Pd, Pt, Rh
- Orange circle: Special metals: Be, Li, Mo, Nb, REE, Sc, Sn, Ta, W, Zr
- Brown circle: Base metals: Al, Co, Cu, Ni, Pb, Zn
- Pink circle: Ferruginous: Cr, Fe, Mn, Ti, V
LEGEND FOR THE INSET MAPS

Sedimentary and or metamorphic, undivided: lighter shades of each are offshore

- C: Cenozoic
- N: Neogene and Quaternary
- P: Paleogene
- K: Cretaceous (often includes Paleogene)
- J: Jurassic (often includes Cretaceous)
- T: Triassic (often includes Jurassic)
- P: Paleozoic (mostly Cambrian to Devonian)
- P: Permian (often includes Triassic)
- C: Carboniferous (often includes Permian)
- D: Devonian (often includes Carboniferous)
- S: Silurian (often includes Devonian)
- O: Ordovician (often includes Silurian)
- C: Cambrian (often includes Ordovician)
- P: Proterozoic (up to Devonian in Yukon and Alaska)
- M: Neoproterozoic (often includes Cambrian)
- N: Mesoproterozoic
- P: Paleoproterozoic
- A: Archean, Archean and Paleoproterozoic

Distinct lithologic suites (with age label prefix): lighter shades of each are offshore

- E: Extrusive; igneous, undivided
- F: Felsic intrusive suites (granite, tonalite)
- G: Gabbro suite (also anorthosite)
- P: Peridotite suite
- X: Impact breccia

Cratons, shield areas

- ASI: Slave Craton: felsic intrusive suites
- AR: Rae Craton: intrusive undivided
- ASB: Hearne Craton: felsic intrusive suites
- AM: Murmansk craton: felsic intrusive suites
- AK: Karelian Craton: felsic intrusive suites
- AN: North Atlantic Craton: intrusive undivided
- A: Anabar shield: sedimentary undivided
MINERAL RESOURCES IN THE ARCTIC

The geological surveys active in the Arctic region have compiled information on the most important mineral deposits north of 60°N - in a database, on a map and in user-appropriate descriptions (for geoscientists in this volume and in a briefer version for general-interest readers). The largest deposits of metals and diamonds on land have been given priority. The briefer version of the description will be published in English, French and Russian.

These products represent the first compilation of information on the most important deposits of the prioritized resource types in the Arctic. The compilation illustrates the importance which the mineral industry has had in the Arctic regions for over a hundred years, but also the priority given to exploration for mineral resources in the region in more recent times. The mineral industry is very important for the northernmost regions of most of the Arctic nations and several deposits north of 70°N (in Canada, Greenland, Norway and Russia) are being mined or developed with a view to mining. Production of certain commodities from mines in the Arctic represents a major part of world production. The results of exploration in both established mining provinces and in new, prospective areas show that there is a considerable potential for new, important discoveries.