GEODE
ore deposits
geodynamics

Ilmenite deposits and their geological environment

With special reference to the
Rogaland Anorthosite Province
Including a geological map at scale 1:75,000
and a CD with a guide to the province

The Rogaland Intrusive Massifs

An Excursion guide

A CD-presentation of the
excursion guidebook and
additional geological map
information, photographs
and SEM images

10 km
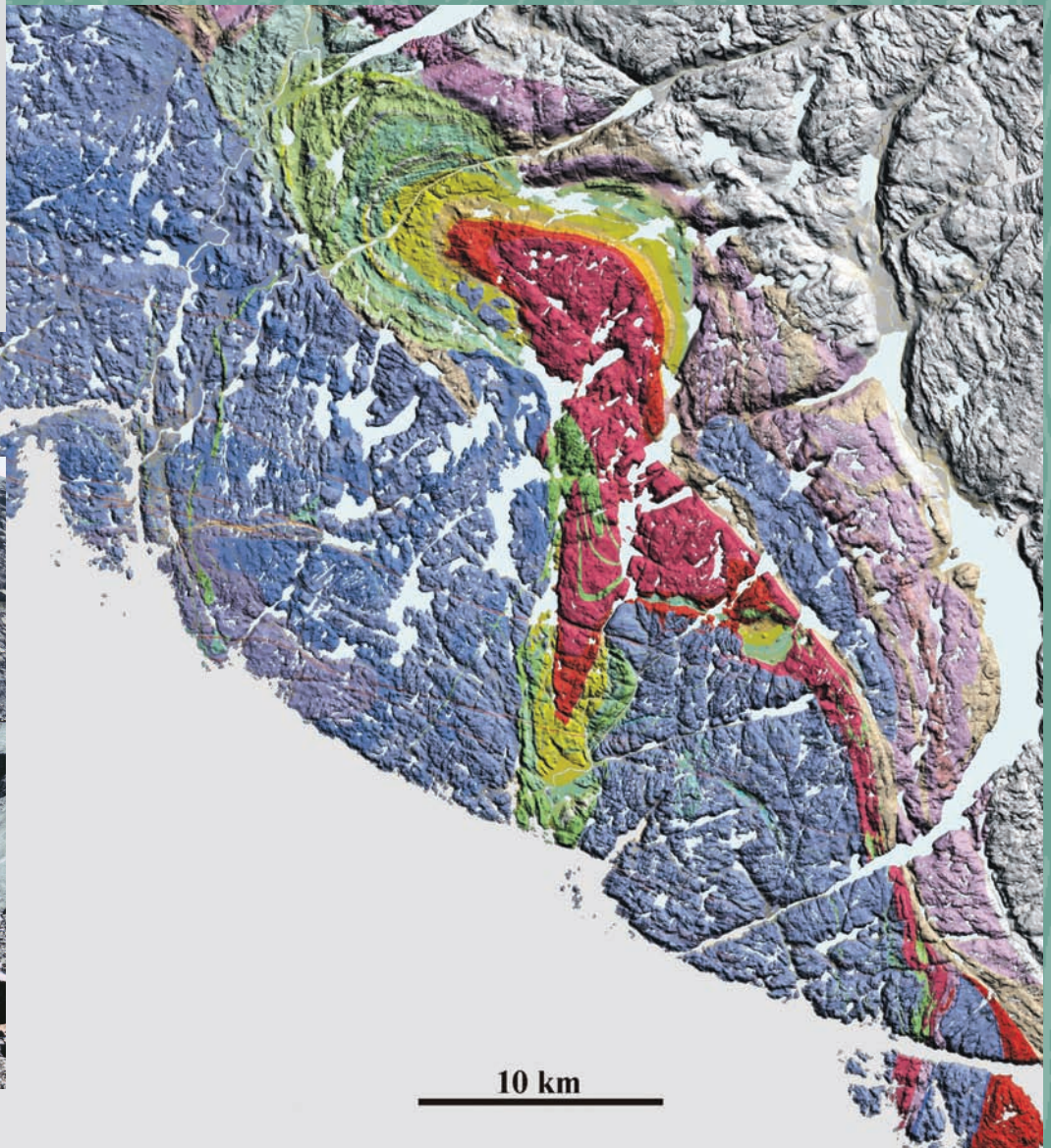
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Cover illustration:
Geological map of the Rogaland Anorthosite Province. The image was produced by John F. Dehls using geological data from Marker et al. (this volume) and topographic data derived from the Norwegian mapping authority (Statens kartverk).
Inset photograph shows the Tellnes open pit in 1964 (photo by Jean-Clair Duchesne).

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Ilmenite deposits and their geological environment

**With special reference to the Rogaland Anorthosite Province
Including a geological map at scale 1:75,000 and a CD with
a guide to the province**

**Edited by:
JEAN-CLAIR DUCHESNE & ARE KORNELIUSSEN**

Trondheim 2003

Images from the Rogaland Anorthosite Province

Modally layered norite within megacyclic unit III of the Bjerkreim-Sokndal layered intrusion at Teksevatnet (loc. 84, Fig. 1 in Schiellerup et al., this volume). Inset photo shows anorthosite xeno-liths in norite near the country rock contact of the Bjerkreim-Sokndal intrusion north of Bjerkreim.

Participants at the GEODE field workshop (2001) visiting the Piggstein anorthosite-norite locality in the Egersund-Ogna anorthosite. The inset photo shows a mega-crystal of orthopyroxene in anorthosite. For further information, see the excursion guidebook (loc. 1.2) available on the enclo-sed CD.

Contents

Enclosures
 - A paper print of the geological map of the Rogaland Anorthosite Province
 - A CD (Guideboook + map + other information interactive)

Preface

The Rogaland Anorthosite Province is famous for its anorthosite massifs, possibly the most accessible in the world with beautiful, nearly continuous outcrops, its numerous ilmenite deposits, which include a world-class deposit (Tellnes), and the largest layered intrusion in Europe (the Bjerkreim-Sokndal body). These geological wonders have been extensively studied for the last 50 years following the impulse of Paul Michot (1902-1999) who compiled the first geological map and suggested an interpretation of the geological evolution. Rogaland, and particularly the Sokndal district, is also famous for its mining activities, not to forget that it was the cradle of the sulphate processing of the ilmenite ore.

The purpose of the present volume is threefold: 1. to present a collection of papers and abstracts given at a symposium on ilmenite deposits and their geological environment; 2. to publish a geological map of the Rogaland Anorthosite Province at scale 1:75,000; 3. to make available a CD with a geological guide-book to the province.

A field-workshop "Ilmenite deposits in the Rogaland Anorthosite Province" was held at Moi, July 8-12, 2001, as part of the European Science Foundation's GEODE project "The Fennoscandian Shield Precambrian Province". The purpose of this workshop was to bring together geoscientists interested in the genesis of the various types of mineral deposits within the overall evolution of the anorthosites (in Rogaland and elsewhere), with particular reference to their geochemical composition. The occurrence of a world-class deposit of ilmenite in the area – the Tellnes mine, run by Titania A/S -– and also the application of new geochemical constraints to the industrial quality of the ore further motivated organising a meeting where the scientific and industrial worlds could interact and compare their experience. Fifty-five participants coming from 14 countries attended the workshop. Forty-four communications (oral and posters) were presented in 2 days. Four days were devoted to field trips. We are particularly happy to present here, besides the abstracts of the communications, a general paper dealing with the gene-

sis of oxide deposits in layered intrusions by G. Cawthorn, and two papers on the deposits of the Ukrainian Shield – D. Gursky, S. Nechaev & A. Bobrov and S. Nechaev & V. Pastukhov, which synthesise for the western scientists a large amount of literature in Russian and Ukrainian.

The geological map at the scale 1:75,000 of the Rogaland Anorthosite Province which is included in this volume has been compiled, revised and locally extended by M. Marker, H. Schiellerup, G. B. Meyer, B. Robins & O. Bolle after the original work of the several groups of geologists who have studied the area. The German group was led by H. Krause from Clausthal. University, the Norwegian group by B. Robins from the University of Bergen, the Danish group by J.R. Wilson from the University of Aarhus, and the Belgian group, following Paul Michot and his son Jean, by J.C. Duchesne from the University of Liège. Some data by the Dutch group of the University of Utrecht have also been included. Ø. Nordgulen (NGU) is particularly thanked for having initiated the publication project as early as 1996.

The excursion guide-book entitled "The Rogaland Intrusive Massifs – an excursion guide" (J.C. Duchesne, editor) NGU report 2001.29, 2001, 137 pp., which is presented on a CD, is the second revised and enlarged version of parts of "Geology of Southernmost Norway, an excursion guide" (C. Maijer & P. Padget, editors) Norges geologiske undersøkelse, Special Publication 1, 1987. The first version was published to illustrate a NATO Advanced Science Institute held at Moi in 1984 on the topic "The deep Proterozoic crust in the North Atlantic Provinces" (A.C. Tobi & J.L.R. Touret, editors) NATO ASI Series, C 158, Reidel, Dordrecht, 1985, 603 pp. The second version was prepared for a meeting held in Rogaland in 1992 on "Magma chambers and processes in anorthosite formation" in the framework of IGCP programme 290 on "Anorthosites and related rocks". Several localities described by Paul Michot in the very first geological guide-book on the region (Norges geologiske undersøkelse, 212g, 1960) are still visited in the present versi-

on. The guide-book, which results from the interaction between several groups of researchers, is divided into two parts: thematic papers introduce the reader to the state of the art on the knowledge of the various units, summarising decades of research up to 2001; then itineraries are described to show salient field exposures of relatively easy access together with petro-geochemical information on the exposed rocks. A map showing the itineraries and the localities on the geological background permits the reader to easily move to the appropriate section of the text.

This volume, which offers an up-to-date synthesis on our knowledge of the Rogaland Anorthosite Province, its geology, petrology and metallogeny, was made possible through the cooperation and interaction between numerous persons with various specialities and interests, from humble students to experienced scientists. The editors wish to thank all of them for their contributions. They also acknowledge D.J. Blundel, Chairman of the Steering Committee of the GEODE programme, A. Bjørlykke, Director of the Geological Survey of Norway (NGU), and R. Hagen, Titania A/S, for encouragement, help and financial support. Thanks also go to NGU secretary Tove Aune for her invaluable help in putting together this volume.

J.C. Duchesne and A. Korneliussen
(23 03 2003)

Short papers

Genesis of magmatic oxide deposits - a view from the Bushveld Complex

Grant Cawthorn

The Bushveld Complex contains 14 and 24 significantly thick layers of chromitite and vanadiferous magnetitite respectively, that can be traced for very many tens of kilometres. Their genesis is still enigmatic. For all models, unresolved problems and contradictory evidence still exist. For chromitite layers, a model involving pressure changes is the most appropriate for explaining the enormous lateral continuity and sharp boundaries. However, the initial $^{87}Sr/^{86}Sr$ of silicate gangue in chromitite layers is anomalously high and indicates mixing with a siliceous component (roof melt?). For the magnetitite layers, extreme gradients of Cr content in vertical sections preclude crystal settling from a large volume of magma, and in situ growth mechanisms may have been operative. The discordant magnetitite pipes may have formed by flushing of a different magma through an unconsolidated crystal pile, feeding into the chamber, with simultaneous dissolution of the host and precipitation of only the liquidus mineral (magnetite). An origin for lenses and bifurcations of silicate gangue in oxide layers is proposed, in which the silicates are considered to be a primary accumulation during a period of non-deposition of oxide, and not due to secondary remobilization or injection processes.

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Introduction

The Bushveld Complex (Fig. 1a) is an extremely large lopolith, covering approximately 65,000 km^2, and up to 8 km thick (see summary by Eales & Cawthorn 1996). Of all layered intrusions, it shows the most extreme range of differentiation (Fig. 1b). The first formed cumulates (containing olivine Fo$_{90}$) suggest crystallization from an almost primitive magma, and the evolved cumulates contain olivine approaching Fo$_0$. The rocks range from dunite and pyroxenite at the base to ferrodiorite at the top, and contain layering (sometimes monomineralic) of olivine, orthopyroxene, plagioclase, chromite, magnetite and apatite.

The Bushveld Complex contains over half the world's chromium and vanadium deposits, contained in numerous oxide-rich layers and, in the case of vanadiferous magnetite, discordant pipes. Over 80% of the world's platinum-group elements also occur here, with about half of this proportion occurring in the uppermost thick chromitite layer. Chromitite layers can be traced for many tens of km laterally in both eastern and western limbs of the intrusion (Fig. 1a). For the lower two packages of layers (referred to as the Lower Group 1-7 and the Middle Group
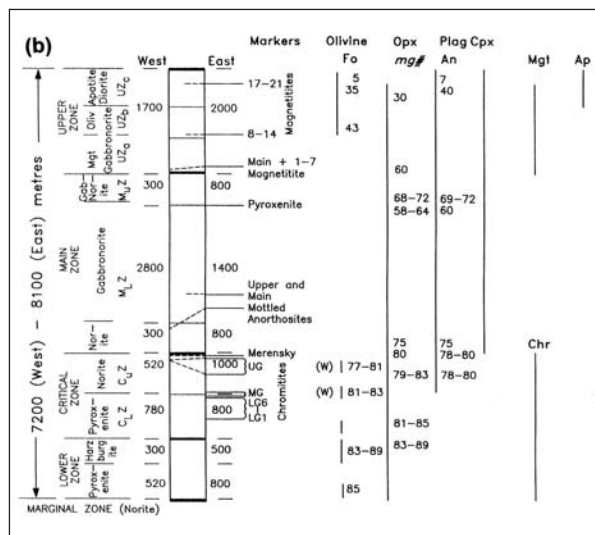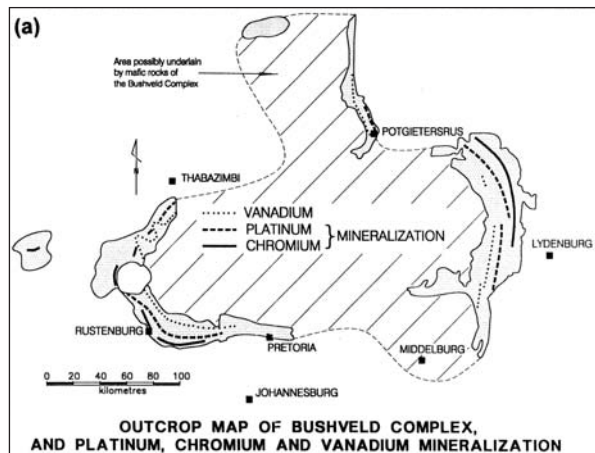
Fig. 1. (a). Simplified map of the Bushveld Complex showing the locations of the chromitite and vanadiferous magnetitite layers. The location of the platinum mineralization is coincident with the uppermost chromitite layer in the intrusion. Structural complications and poor outcrop in the northern part of the eastern limb make exposure of the magnetitite layers discontinuous. (b). Simplified stratigraphic section through the Bushveld Complex, showing the presence of cumulus minerals and their compositions. Sources of data are summarized in Eales & Cawthorn (1996).

1-4) thicknesses remain constant for long distances, but abrupt changes in thickness may occur (Schurmann et al. 1999), usually related to structural features (either folds in the floor or lineaments, that divide the lowest succession into smaller compartments). The layers in the topmost package (the Upper Group 1-3) are remarkably uniform in thickness for over 100 km in each limb. Chemical similarity is also very uniform even across the entire intrusion (Cawthorn & Webb 2001). Chromitite layers typically occur at the base of cyclic units in the Critical Zone, where such can be identified by changes in the silicate mineralogy. Cycles typically consist of a basal chromitite, pyroxenite or harzburgite, norite and anorthosite. Contacts of the chromitite are generally sharp, but the Lower Group 6 chromitite has gradational contacts top and bottom over about 20 cm. These major layers range from tens of cm to 1 m in thickness, but many thinner layers (some even less 1 cm) also occur, and can be equally laterally extensive. They contain from 60-90% chromite, cemented by oikocrysts of pyroxene and/or plagioclase.

They range in thickness from a few cm up to tens of cm, with two exceptions. The Main Magnetitite Layer (fourth from the bottom) is 2 m thick, and the uppermost layer is up to 6 m thick, but contains numerous autoliths and individual grains of silicate material. The oxide content can range up to 95% titanomagnetite, with interstitial altered plagioclase.

A large number of erratically distributed, discordant pipes, trending orthogonally to the layering, are known (Scoon & Mitchell 1994). They range from almost pure magnetitite to coarse-grained, magnetite-rich melagabbro. The largest is over 300 m across and has been drilled to a depth of 300 m, but the total vertical continuity of these pipes cannot be assessed.

Chromitite layering

The various models that have been proposed in the formation of chromitite layers include (in very approximate chronological order):
1. crystal settling and sorting (Sampson 1932);
2. immiscibility of a Cr-rich liquid (Sampson 1932, McDonald 1967);
3. increase in oxygen fugacity (Cameron & Desborough 1969);
4. contamination by siliceous components (Irvine 1975);
5. mixing between resident and new magma (Irvine 1977);
6. lateral growth within a stratified magma column (Irvine et al. 1983);
7. pressure change (Osborn 1980);
8. injection of a chromite-porphyritic magma (Eales 2000).

Each model explains certain aspects of the geology or magmatic behaviour, but equally valid reasons can be presented as to why each process can be questioned.

Fig. 2. Phase diagrams in the system forsterite - quartz - chromite. (a). Magma contamination model after Irvine (1975). (b). Magma mixing model after Irvine (1977).

The magnetite layers are even more laterally extensive and uniform in thickness than the chromitite layers, being traced for over 100 km in each limb. Even such features as the 'feldspar parting', a 20 cm-wide zone that contains about 15% cumulus plagioclase, exactly in the middle of the 2 m-thick Main Magnetitite Layer, are universally present. Almost all layers in the eastern limb have sharp basal contacts with anorthosite, and gradational top contacts, although one layer displays the reverse relationships (Molyneux 1974).

A quantitative assessment of the envisaged processes, specifically a mass-balance evaluation, presents problems. Chromitites contain about 45% Cr_2O_3 or 300,000 ppm Cr. Mafic magmas may contain between 100 and 1,000 ppm Cr. Hence, even if every atom of Cr could be extracted from a magma, between 300 and 3,000 units of magma would be needed to make one unit of chromitite. For example, allowing for the density of chromitite being 1.5 times that of a magma, a one metre-thick layer of chromitite would need between 300 and

3,000 metres of magma. However, the supposition that some process could remove every ion of Cr from the magma is obviously invalid, and so very much greater thicknesses of magma need to be involved. It should be recalled also that most layers of chromitite can be traced for many 10s if not 100 km in the east and west Bushveld, and could extend for 300 km if the principle of connectivity of the two limbs is accepted (Cawthorn & Webb 2001). Processes that can operate on such huge volumes of magma need to be evaluated.

If crystal settling has taken place, then Stokes Law, with or without the complexities of non-Newtonian behaviour of magma, needs to be applied. Although chromite has a high density, its small grain size relative to pyroxene and olivine makes is unlikely to sink more rapidly than the mafic minerals. Hence, producing a chromitite layer at the base of a cyclic unit, overlain by ultramafic layers, is not consistent with Stokes Law. Further, with reference to Fig. 2, the olivine to chromite ratio along the cotectic ranges from 98 to 2 (for point A) to about 99.6 to 0.4 (at point B). To make a 1 m-thick chromitite layer from a magma at the cotectic would require the implausible situation of maintaining 50 to 250 m of olivine crystals in suspension. Hence, the question in producing a chromitite layer can be defined as: why were there no silicate minerals growing and sinking while the chromite was accumulating?

The concept of Cr-rich immiscible liquids was developed by Sampson (1932) partly to explain the bifurcations exhibited by chromitite layers, as seen for example at the Dwars River locality (see below), where it was suggested that a dense, Cr-rich liquid was injected downwards into a plagioclase crystal mush. However, no experimental duplication of liquid immiscibility for Cr-rich liquids has ever been achieved.

The stability of oxide phases depends upon the ferric ion activity in the magma, and hence an increase in fO_2 would cause chromite (and magnetite) precipitation. Upward percolation of fluid has been suggested as the source of the oxygen, but whether such percolation could occur over such laterally extensive distances to produce uniformly thick layers of chromitite (and magnetitite) seems dubious.

Also, fO_2 changes would not explain the common pattern of cyclic units in which anorthosite occurred below and pyroxenite above chromitite layers.

Another model for chromitite formation is that of magma mixing, and will be discussed in some detail here. In its earliest form, Irvine (1975) suggested that contamination of magma (composition A in Fig. 2a) by a siliceous component (F) could have lowered the solubility limit of Cr in a mixed magma (M), and so induced chromite separation. In support of this model, the initial $^{87}Sr/^{86}Sr$ values for interstitial silicate minerals in several chromitite layers have recently been reported (Schoenberg et al. 1999). Their results are shown in Fig. 3. The high ratio observed in the chromitite layer itself compared to its footwall suggests that massive degrees of crustal contamination occurred. However, what is more remarkable is that the hangingwall rocks to the layers have a ratio that does not show evidence of this contamination. The implication is that the chromitite formed from a large volume of contaminated magma, perhaps generated by mixing with molten felsic roof rocks, but that the overlying pyroxenite formed from a magma that did not show this contamination effect.

A further variation of the magma-mixing model was introduced by Irvine (1977), in which mixing between differentiated magma (B or C in Fig. 2b) and primitive magma (A) caused the mixed magma (M or M^1) to lie in the chromite stability field. Excess chromite formed, producing the layers. These models have the advantage over the crystal settling and sorting model, in that it explains why the magma lies in the chromite stability field, and so does not crystallize any mafic minerals. The effect on the Cr budget of the magma as this process occurs can be seen in Fig. 4. The degree of supersaturation with respect to Cr of the mixed magma increases as the difference in MgO contents of the mixing magmas also increases. It was suggested (Murck & Campbell 1986) that near the base of large layered intrusions chromitite layers are not developed, because the resident magma is not very different from the added, primitive magma. However, even for magmas that are very different, the extent of supersaturation may not exceed 100 ppm Cr
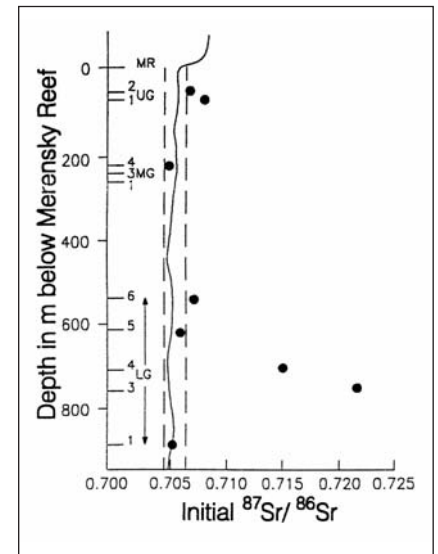
Fig. 3. Plot of the initial $^{87}Sr/^{86}Sr$ in the several chromitite layers compared to the typical ratio throughout the Critical Zone (from Schoenberg et al. 1999). The anomalously high value in the chromitite layer suggests a very large degree of contamination.

Fig. 4. Plot of Cr versus temperature of magma saturated with respect to chromite (Murck & Campbell 1986). Mixing line DP shows the oversaturation in chromite induced by mixing. M-M1 shows the excess Cr available to produce chromite.

Fig. 5. Plot of mg# in orthopyroxene across the LG6 chromitite layer. Data are taken from Cameron (1980). Open circles refer to cumulus orthopyroxene, solid circles to interstitial orthopyroxene in chromitite layers in which there has been re-equilibration between the phases in terms of Mg and Fe.

(i.e. difference between M and M¹ in Fig. 4). If this figure represents the extractable Cr as chromite prior to silicate saturation, a one m-thick chromitite layer would require the processing of 4,500 metres of magma.

The mineral compositions can also give a clue as to the plausibility of this process. In the case of the Bushveld Complex, the parent magma (point P in Fig. 4) containing 13% MgO would crystallize orthopyroxene with an mg# of 90. The more evolved magma crystallizing plagioclase in the Upper Critical Zone (point D in Fig. 4) containing about 7% MgO would crystallize orthopyroxene mg# 82. Mixing of these two magmas ought to have produced a mixed magma that crystallized orthopyroxene with mg# about 86. Thus, magma addition ought to be recognizable by reversals in mineral composition. Inspection of the mg# below and above the economically most important and thickest layer - Lower Group 6 chromitite layer shows that the composition actually shows a decrease in mg#, not a reversal (Fig. 5). The data in Fig. 5 include cumulus orthopyroxene in the footwall and hangingwall, and also intercumulus orthopyroxene in the chromitite layer. Within the chromitite layer there is an increase in mg#. However, this increase is

not a primary magmatic effect indicating addition of primitive magma. It is a secondary effect, due to re-equilibration between orthopyroxene and chromite, causing the orthopyroxene grains to increase in mg#. The data of Cameron (1980, 1982) show that, across all chromitite layers, there is a change in mg# in orthopyroxene of less than 1 (with an increase above the layer being less common than a decrease), changes that do not support the concept of addition and mixing of undifferentiated magma.

There is a second aspect that also needs to be considered, namely the small-scale field relations. A very few chromitite layers have gradational contacts, but most have very sharp contacts (Fig. 6a). Any mixing process would have to be instantaneous in order to produce such a sharp contact. The mixing of such huge thicknesses of magma over such large horizontal distances so effectively seems extremely difficult to envisage.

The lateral growth model, proposed by Irvine et al. (1983) requires that crystallization only occur at the margins of a stratified magma column. This model is a special case of in situ growth. If in situ growth is the mechanism by which these layers form, it becomes necessary for Cr to diffuse or be transferred to the base or edge of the magma column to sustain chromite growth. Transport of such huge quantities of Cr as would be needed to produce a one m-thick layer of chromitite seems implausible.

Yet another model was originally proposed by Osborn (1980) and has been elaborated upon by Lipin (1993). In this model, an increase in pressure occurred in the magma chamber to initiate chromite precipitation. The process can be seen in terms of phase diagrams for the same system but at different pressures (Fig. 7). The stability field of chromite increases with pressure relative to pyroxene and plagioclase. A magma may lie at the three-phase invariant point (point C) at low pressure, but an increase in pressure would cause the same magma to lie in the chromite field. Demonstration of pressure changes in magma chambers is hard to prove. However, the fountaining of basic lava, followed by draining of lava lakes over a period of days in active volcanoes demon-

Fig. 6. Nature of contacts and detailed field relations of oxide-dominant layers. (a). Extremely sharp contacts of chromitite layers in anorthosite. The sharp, but somewhat cuspate, upper contact is a common feature. (b). Typical contacts of magnetitite layer, showing sharp base and gradational top, in both cases with anorthosite. (c). Detail of lenses and bifurcations within chromitite layers from Dwars River locality of Upper Group 1 chromitite layer.

strates that pressure can fluctuate at least under volcanoes (Pollard et al. 1983). The physical mechanism operative within a large magma chamber is harder to postulate. However, if magma is added to the chamber, the sides must expand or the vertical height must increase. In the latter case, the pressure at the base of the chamber must also increase. Alternatively, if the extra load of added magma causes the floor to collapse (Carr et al. 1994), the magma at the base would be subjected to an increase in pressure. If the magma chamber were sealed, an attempt to add magma would result in a transmission of a pressure wave without actual magma addition since liquids are relatively incompressible. Whether a sufficient column of magma could be affected remains speculative, again emphasizing the mass balance problem. However, pressure changes can be transmitted everywhere essentially instantaneously within the chamber, and so crystal-sharp boundaries to layers and lateral uniformity could be envisaged.

It has also been suggested that added magma carried a suspension of chromite crystals, which produced a chromitite layer (Eales 2001). However, given the density of chromite it is hard to envisage the suspended chromite being uniformly deposited laterally, at great distances from proposed feeder sites. This model also begs the question as to what process in the deeper magma chamber or source area permitted the formation of such a huge mass of chromite. If it could happen in a deep magma chamber, why could not the same process occur in the Bushveld chamber?

In conclusion, it appears that either there are reasons why the apparent contradictory evidence presented here for the currently proposed models can be ignored, or that some other processes need to be identified to explain chromitite layering.

Magnetitites

Whereas most occurrences of magnetitites are as layers in the Upper Zone of the Bushveld Complex, pipes or almost pure magnetite up to 300 m in diameter and rarely narrow dykes, are also recorded from both the Upper and the Main Zones. Similar models to those for chromitite formation have also been applied to the generation of magnetitite layers. Some of

the models may be questioned on similar grounds to those listed above. Other processes and problems are summarised below.

Unlike the experimental rebuttal of the existence of Cr-rich immiscible liquids, iron-rich immiscible liquids have been synthesized and are petrographically identifiable (Philpotts & Doyle 1983). However, applicability to the origin of oxide layering is questionable. If a dense iron-rich liquid did form, it ought to sink into the underlying plagioclase crystal much, and not form a regular planar base (Fig. 6b). Further, if this dense liquid did accumulate into a planar body, it would not be possible for plagioclase crystals to sink into it to form a gradational upper contact (Fig. 6b), that is characteristic of most layers. Further, the composition (Table 1) of such immiscible liquids has been shown by Philpotts & Doyle (1983) to contain only 30% FeO (total). It is hardly of a magnetitite composition, and so still requires a large proportion of silicate material to be removed from it to become a
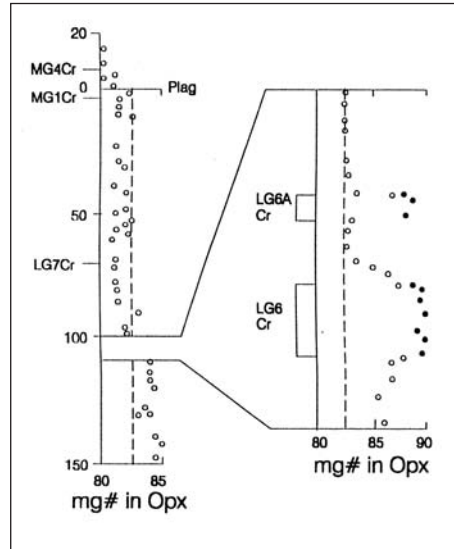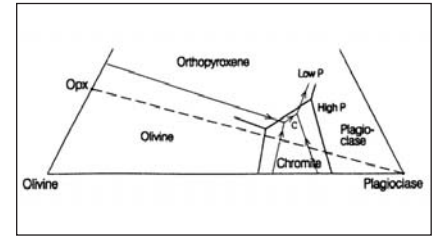

Fig. 7. Schematic phase relations in the natural system olivine-plagioclase-quartz at two different pressures, illustrating the enlargement of the chromite stability field with increasing pressure (modified from Osborn 1980).

Table 1. Compositions of immiscible liquids (natural and experimental examples).

	Natural Example			Experimental Analogue		
	Pre-immiscibility	Si-Rich Liquid	Fe-Rich Liquid	Pre-immiscibility	Si-Rich Liquid	Fe-Rich Liquid
SiO_2	79.1	73.3	35.6	58.8	72.3	44.3
TiO_2	0.1	0.2	3.8	1.7	0.3	2.1
Al_2O_3	10.3	11.9	2.2	10.2	11.7	7.3
FeO(T)	1.7	2.6	33.7	16.6	6.7	28.6
MgO	0.2	0.0	0.5	0.6	0.0	1.0
CaO	0.8	1.4	10.1	5.4	3.8	9.5
Na_2O	2.6	3.0	0.0	2.9	2.9	1.0
P_2O_5	0.0	0.1	3.0	0.7	0.4	0.9

Note: the column headed 'pre-immiscibility' indicates the composition of the liquid at a temperature immediately above that at which liquid immiscibility appeared (Philpotts & Doyle, 1983).

pure magnetitite layer. Finally, the composition of the silicate magma at which immiscibility occurs is highly evolved, containing little MgO and so being incapable of forming the composition of mafic minerals observed in association with magnetitite layers. For example, the orthopyroxene at the level of the lowest magnetitite layers has an mg# of 60 (Fig. 1b). Iron-rich immiscible magmas also strongly partition phosphorus relative to the silicate magma (see Table 1). It is for this reason that nelsonites, which are
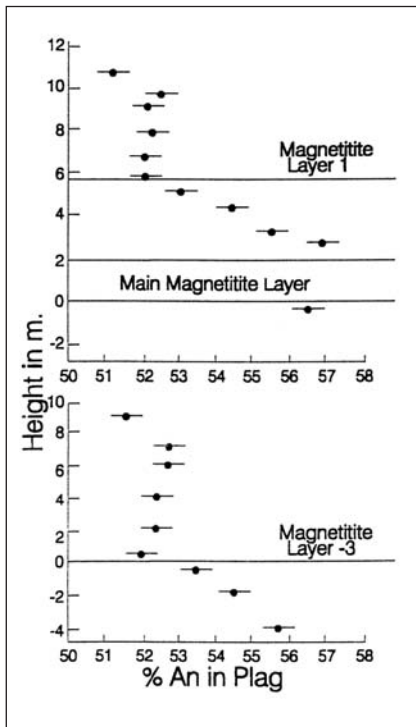
Fig. 8. Plot of An content in plagioclase versus height across three magnetitite layers near the base of the Upper Zone, showing absence of reversals. Bars are range of core compositions from several grains. There is a vertical gap of about 30 m between these two sections, and a reversal in An content occurs in this interval. (Author's unpublished data.)
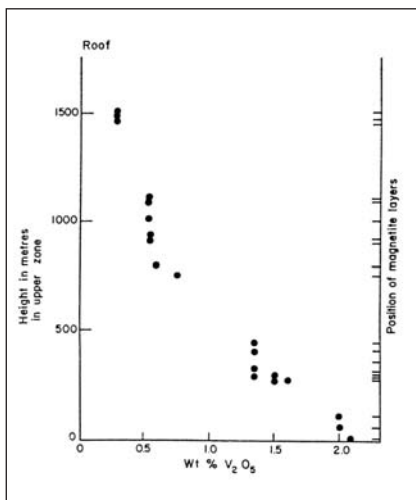
Fig. 9. Plot of V_2O_5 content in magnetitite layers versus height in the Upper Zone, Bushveld Complex (from Molyneux 1974).

oxide-apatite rocks, have been considered to result from liquid immiscibility. This hypothesis has also been applied to the magnetitite layers of the Bushveld Complex by Reynolds (1985). However, it should be noted that the lower layers of magnetitite contain no apatite at all, and that the apatite-bearing layers only appear above the level at which apatite becomes a cumulus mineral in the associated silicate layers (Cawthorn & Walsh 1988). Thus, the appearance of apatite with magnetite does not suggest liquid immiscibility, but simply the normal progression of magmatic fractionation.

As with chromitite layer contacts, the bases of magnetitite layers are extremely sharp (Fig. 6b), although upper contacts are often gradational (Fig. 6b). The efficacy of magma mixing to produce such abrupt contacts is questionable. The Upper Zone magma displays a remarkably constant and anomalously high initial Sr isotope ratio (Kruger et al. 1987). (Analogous Sr isotope data to those presented in Fig. 3 for the interstitial components of chromitite layers have not been compiled for magnetitites, because the plagioclase shows significant alteration in surface and shallow borecore samples, but such data would be critical in view of the implications for chromitite genesis.) The Upper Zone magma is inferred to be the mixture from several partially crystallized magmas, all with different isotopic ratios, that produced the preceding zones. It is unlikely that a new magma could have had exactly the same ratio as this complexly mixed resident magma. Hence, the constancy of initial ratio throughout the entire Upper Zone argues against addition of significant volumes of distinct magma. Analyses of plagioclase in the footwall and hangingwall of several layers, shown in Fig. 8, suggest that they are related by differentiation and not addition of magma. Since these rocks are quite evolved (An_{55-60}) addition of undifferentiated magma ought to produce significant reversals in composition, but they are not observed. Reversals in the An content of plagioclase have been identified in the Upper Zone (Christian Tegner, unpublished research), but they are not related to the presence of magnetitite layers. For example, a reversal occurs *between* magnetitite layers, not across a magnetitite layer, in Fig. 8.

There is a fairly regular upward decrease in the V content of the magnetite (Fig. 9). Again, if undifferentiated magma, with high V content, had been added to the system, reversals or irregular trends might be expected.

Cr is highly compatible in magnetite (the partition coefficient is in the order of hundreds) and is a sensitive indicator of fractionation. Its abundance in magnetite (Fig. 10) decreases extremely rapidly in short, vertical sections through magnetitite layers, and also shows abrupt reversals even within layers. The rate of depletion is not consistent with crystal settling from a large volume of magma, which would have produced either homogeneous layers or subdued depletion profiles in Cr. The rapid depletion profiles were interpreted by Cawthorn & McCarthy (1980) to be the result of in situ growth. This rapid depletion in the Cr content in the magma seems to contrast with the mass balance calculations present above for chromitite layers, where enormous volumes of magma need to be processed. Despite their apparent similarities, it seems that chromitite and magnetitite layering may not originate by identical mechanisms. The occurrence of reversals in Cr content within layers was attributed to the effect of convection cells sweeping away magma partially depleted in Cr in the boundary zone of crystallization and replenishing it with Cr-undepleted magma. A further consequence of the preservation of such sharp reversals relates to hypotheses about infiltration metasomatism or compositional convection. Significant mineral re-equilibration during vertical migration of residual liquid from below would smooth out such sharp reversals or produce a zone within the magnetitite layer in which the Cr content gradually increased. Such patterns have not been observed.

Finally, it is possibly significant that the footwall to every magnetitite layer in the eastern limb is an anorthosite (Molyneux 1974). Random addition of magma would not be expected to be presaged by an anorthosite below the magnetitite layer. A complete solution for the genesis of magnetitite layers needs to take such observations into consideration. The succession of anorthosite underlying and frequently overlying magnetitite layers begs the ques-

tion as to why no pyroxene is present. The deeper footwall is gabbroic, indicating a magma lying on the plagioclase-pyroxene cotectic. If plagioclase could form and sink, so should pyroxene, hence its absence means that the magma has moved out of the stability field of pyroxene. The hypothesis involving an increase in pressure within the magma chamber, which might apply to chromitite layers, may also have application here. An increase in pressure might drive the magma into the stability field of magnetite to produce a magnetitite layer. However, pressure decrease might have a different effect. It is well established that the thermal stability of mafic minerals and plagioclase increase at different rates with increasing pressure (Fig. 11). Thus, a liquid initially saturated in plagioclase and pyroxene would crystallize only plagioclase if there were a reduction in pressure in the system. In this way, the apparent non-crystallization of pyroxene during the formation of anorthosite in the footwall to magnetitite layers might be explained by pressure reduction (Naslund & McBirney 1996).

The discordant, pipe-like bodies of magnetitite cannot be attributed to pressure changes, and different processes must be involved. They cannot be attributed to immiscible liquids (Scoon & Mitchell 1994) for the same reasons as discussed above. Iron-rich liquids only contain about 30% FeO (total) and so remaining silicate component still has to be removed. Also the immiscible liquids are highly evolved (Table 1) and would not be able to produce magnetite with high V contents as are found in the ore body at Kennedy's Vale in the eastern limb (Willemse 1969).

Discordant silicate-rich pipes in the eastern limb have been modeled as the result of infiltration of a magma that selectively replaced pre-existing non-liquidus minerals with liquidus minerals (Cawthorn et al. 2001). This process is essentially the same as that proposed by Keleman et al. (1992) for the formation of discordant refractory veins of harzburgite and dunite in mantle rocks. In this model, an injecting magma that is out of equilibrium with the host rock, either because it is hotter or chemically undersaturated in some of the minerals in the host rock, replaced existing host rock by precipitating the liquidus mineral of this added magma.

Location of these bodies may have a structural control. For example, the Kennedy's Vale pipe occurs within the Steelpoort fault system (Willemse 1969). In terms of the genesis of oxide deposits, it should be noted that certain ilmenite vein deposits have been similarly interpreted by Duchesne (1996), who argued that pure monominerallic bodies did not have to represent a liquid of that composition, but that only the liquidus mineral separated during extensive through-flow of magma in an open system.

However, if this model is applied to the magnetitite pipes it implies the existence of new magma being added to the complex that had magnetite on the liquidus. Such suggestions would support the concept of addition of magma to make magnetitite layers, which was questioned above on Sr isotopic considerations.

Silicate lenses and bifurcations in oxide layers

Within chromitite and magnetitite layers there occur numerous lenses and bifurcations (Fig. 6c) containing silicate-dominant material. Scale may range from a few cm upwards. A bifurcation may simply be a lens that is of such a size that the entire lens shape is not apparent on the scale of the exposure. The most famous bifurcations can be seen at the Dwars River locality of the Upper Group 1 (UG1) chromitite layer in the eastern limb. In fact, in all exposures the UG1 displays such features (Lee 1981). Apart from some rare calc-silicate xenoliths derived from the metasedimentary remnants from the roof of the intrusion, the silicate partings in oxide-rich layers have the same mineralogy as adjacent layered rock types. Irregularly shaped fragments also occur that are probably autolithic. However, the lenses and bifurcations discussed here have a length to breadth ratio that gives them the shapes of extremely thin disks, which may be only cm thick. It is suggested that they are too thin, and hence fragile (Fig. 6c), to be physically transported fragments.

This author's preference for the origin of these lenses is that they are primary accumulations of silicate minerals that did not develop into laterally extensive layers. The process of formation is shown in Fig. 12.
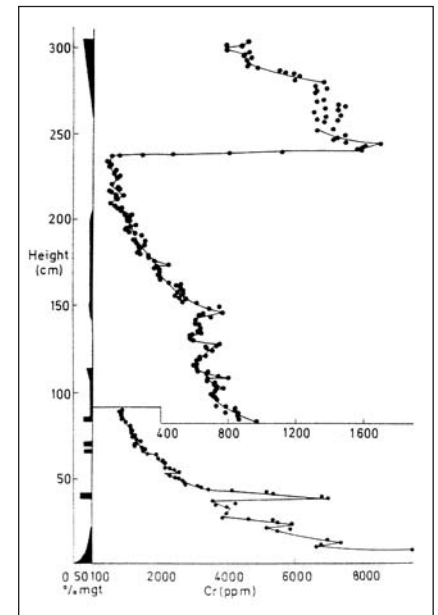
Fig. 10. Plot of Cr in magnetite separates, analysed by XRF versus height from the Main Magnetitite Layer (from Cawthorn & McCarthy 1980).
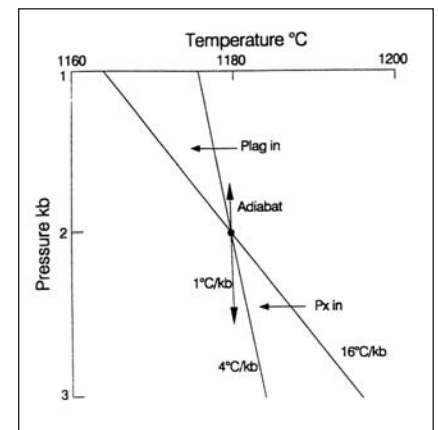
Fig. 11. Plot of thermal stability of plagioclase and pyroxene in a typical basic magma as a function of pressure.
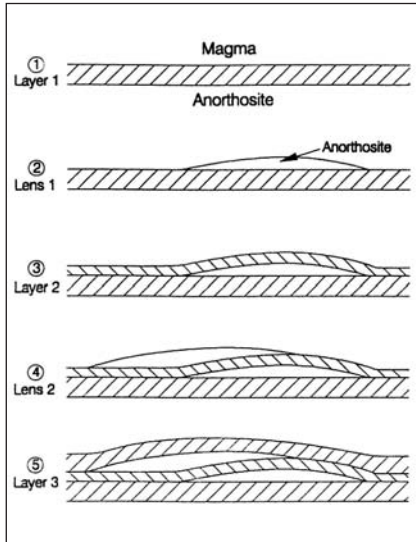
Fig. 12. Schematic model for the formation of lenses and bifurcations in oxide layers. See text for explanation.

This diagram is constructed to explain a lens of anorthosite within bifurcating chromitite layers, but could also apply to magnetitite layers, and also for lenses of any silicate assemblage. A thin layer of oxide (layer 1) formed on top of a continuous layer of anorthosite (stage 1, Fig. 12). The exact process of oxide formation is not important to this model. This continuous layer of oxide was then followed by the development of a partial layer of silicate. Two different models may be envisaged. If layering forms by in situ growth, it is possible that initiation of silicate formation will not occur everywhere on the oxide surface instantaneously (Cawthorn 1994). Instead, growth nodes may form at random spacing on the surface (only one node is shown in stage 2, Fig. 12). If growth continues and the nodes coalesce into an extensive layer, their original existence will not be identifiable. However, if growth is terminated before coalescence by another period of oxide formation, these nodes will be trapped as lenses within a doublet layer of oxide (stage 3, which shows the formation of layer 2 of oxide, Fig. 12). Alternatively, localized patches of silicate mineral accumulation may result from convective activity. If low-density plagioclase accumulates at the base of the chamber by being carried down on descending limbs of convection cells, grains may accumulate only at the base of such descending columns, and not where the cell is flowing horizontally, or ascending. Thus, discontinuous patches of plagioclase accumulation will result. Again, accumulation is terminated by initiation of the second oxide layer (stage 3, Fig. 12), and a lens of anorthosite results. These lenses could be extremely thin relative to their lateral extent. A repetition of oxide termination, localized plagioclase accumulation, followed by oxide layer formation is shown as stages 4 and 5 (Fig. 12), and produces the second anorthosite lens and the third oxide layer.

The consequences of an evolution as shown in Fig. 12 are:
1. Whereas their shape argues against physical transportation, extremely thin (and hence fragile) lenses of silicate material can form and survive as primary precipitates within oxide layers.
2. If the lens is larger than the scale of outcrop, the lens might be seen as a bifurcation only.

3. If two such lenses occur at different stratigraphic levels within the oxide layer, a bifurcation within an oxide layer will be seen to continue laterally until it joins another chromitite layer (above or below). Thus, no terminations of oxide layers within anorthosite bodies should occur, consistent with field observations.
4. Apparently thick oxide layers are not single layers, but the result of repeated formation of thin oxide layers. No silicate material accumulated between the layers to provide a marker that a nonconformity exists within the oxide layer.
5. Reversals in chemical trends (such as the Cr in magnetitite) are in agreement with this hypothesis of multiple discrete oxide layers superposed without intervening silicate.

Numerous other features are evident in such outcrops as the Dwars River locality, and the above discussion does not specifically cover these other processes, but the principle of lenses and bifurcations can be explained as a primary accumulation process rather than as a secondary remobilization event.

Nucleation and supercooling considerations

The previous discussions have all focused on processes that occur during slow cooling and under equilibrium conditions. General models exist for layering that appeal to non-nucleation, oscillatory nucleation, supercooling and other disequilibrium processes, as reviewed by Naslund & McBirney (1996). The fact that magmas are dynamic systems and not always at equilibrium is inescapable. However, in very slowly cooling magma chambers the extent of deviation from ideality may be limited. Gibb (1974) studied experimentally the extent of supercooling prior to plagioclase nucleation. In rapidly cooling systems, supercooling was significant. However, he pointed out that in slowly cooled intrusions (he used the Stillwater intrusion for his example) the rate of cooling is so slow that supercooling and supersaturation would have been very restricted. For example, Cawthorn & Walraven (1998) showed that in the Critical Zone of the Bushveld Complex the rate of cooling was about 1°C every 200 years, severely restricting the opportunity for

supersaturation and non-equilibrium processes. However, the sharpness of oxide layer boundaries, and the appearance or disappearance of associated silicate minerals indicates that some fundamental change had to have occurred in the chamber to produce these oxide layers. For example, magma mixing could cause superheating or supercooling depending upon the detailed topology of the liquidus of the two magma and their intermediate compositions. Pressure changes would also affect the instantaneous liquidus temperature. Thus, either of these processes might lead to non-equilibrium effects, but purely internal processes related to cooling within a single magmatic system cannot provide the answer to oxide layering.

Summary

The Bushveld Complex contains many economically important layers of chromitite and vanadiferous magnetitite. Their lateral extent and chemical uniformity demand processes that can operate instantaneously over enormous distances. Numerous studies yield conflicting interpretations on the origin of these layers. The chromitite layers often form the basal layer of a cyclic unit; magnetitites are usually underlain and overlain by anorthosite within a gabbroic sequence. Models that do not encompass processes producing the distinctive silicate rock association may be too myopic. Each model explains some of the field or geochemical features observed, but contradictory information can be presented against every hypothesis. It is even inferred here that the two types of oxide layering may have different geneses, based on a study of the contrasting behaviour of Cr in both settings.

Acknowledgements

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Titanium deposits in Ukraine focused on the Proterozoic anorthosite-hosted massifs

Dmitry Gursky, Sergey Nechaev & Alexander Bobrov

The largest titanium resources of Ukraine are related to anorthosite parts of large Proterozoic (1.8 Ga) composite plutons; these contain a variety of Fe-Ti deposits with significant differences in ore geological and mineralogical characteristics due to local geological variations. Some well-documented deposits, such as Stremigorod and Krapivnia within the Korosten pluton, and Nosachev within the Korsun-Novomirgorod pluton, contain very large resources of ilmenite ± titanomagnetite ± apatite, although these are not economic at present. Of significant economic importance are residual deposits related to Mesozoic weathering of Ti-bearing gabbro-anorthosite massives and ancient placer deposits, which are presently being exploited and contribute the basis for a Ukrainian titanium industry. Main economic components in these deposits are ilmenite ± rutile ± leucoxene; in some of these deposits other minerals such as apatite, zircon, kyanite, sillimanite, staurolite and diamonds represents valuable by-products.

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Introduction

Titanium deposits and prospects in Ukraine are associated with Proterozoic and Palaeozoic magmatic basic rocks, Mesozoic and Cenozoic weathering crusts of basic rocks and alluvial placers, and Cenozoic littoral placers. Reserves and resources of Ti-ores are a stable basis for the Ukrainian titanium industry. The basis for the latter are the exogenic deposits of Ti-ores, whereas endogenic ores are not mined at present. Nevertheless the main Ti resources are hosted by gabbro-anorthosite massifs.

The Ukrainian Ti-ore province, as described by Malyshev (1957), coincides spatially with the outlines of the Ukrainian Shield (USH) Precambrian complexes and their Mesozoic and Cenozoic cover. Later, the borders of the province were enlarged

into the Dnieper-Donets aulacogen (Fig.1).

During the time of the former USSR, a large Ti-mineral industry was created in the Ukraine. The total share of the Ukrainian TiO_2 industrial reserves in the USSR was 42%, including more than 90% in the European part of the country. The basis of the industry includes more than 40 titanium deposits, 11 of which are explored in detail. Only 5 of them, of exogenic origin, are being exploited. The other Ti-deposits are reserves. TiO_2 production in the Ukraine is shown in Table 1.

Exogenic deposits

The exogenic Ti-ores are of residual and placer types. Residual deposits are spatially related to Mesozoic weathering of Ti-

Table 1. Overall production of ilmenite and rutile concentrates in Ukraine during the period 1988-1998 (thousands of tons recalculated as TiO_2, after Belous 1998, and World Mineral Statistics 2000).

Year of production	1988	1989	1990	1991	1992	1994	1995	1996	1997	1998
Ilmenite concentrate	263.0	249.3	229.6	217.2	221.4	530	359	250	250	250
Rutile concentrate	118.2	130.4	123.6	108.0	108.7	80	112	50	50	50
Total	381.2	379.7	353.2	325.2	330.1	610	471	300	300	300
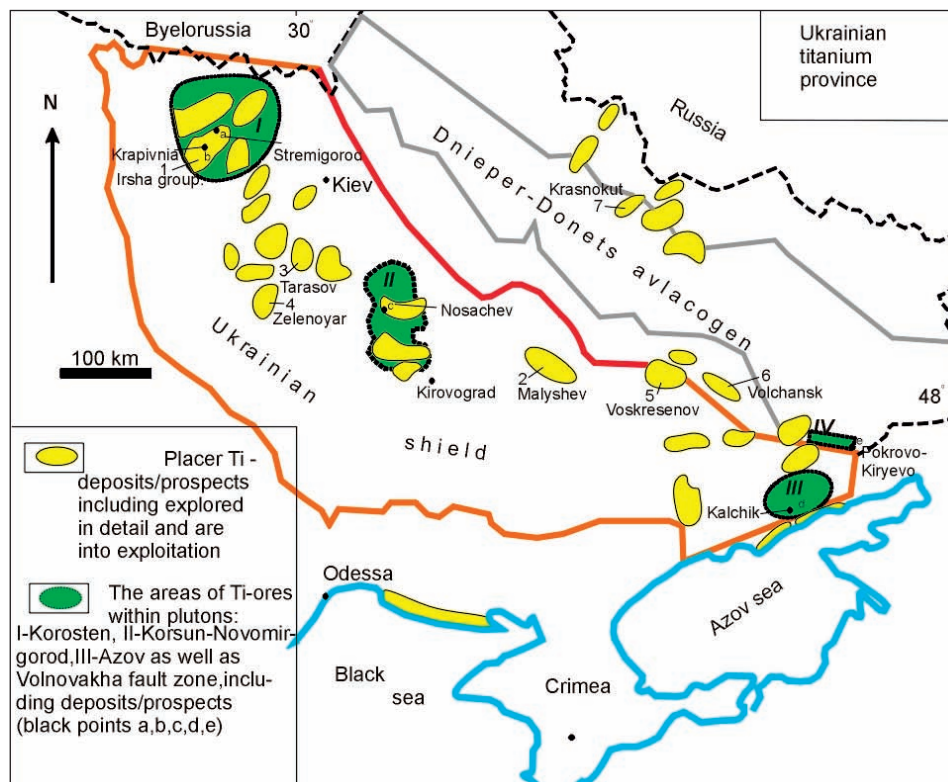
Fig. 1. Sketch map showing the location of titanium deposits and prospects within the Ukrainian Ti-province (after Bochai et al. 1998)

Table 2. The principal geological-industrial types of exogenic titanium deposits (after Belous 1998).

ilmenite and apatite, respectively. Placer deposits are the main industrial Ti-ores in Ukraine, particularly complex Ti-Zr littoral placers, although continental, alluvial, ilmenite placers are also important. The latter are developed to the northwest of the USH and are related to intense Late Jurassic-Early Cretaceous weathering and erosion. There are both ancient and modern continental placers. All of the ancient placers are buried. These are the largest and richest deposits: ore-bodies are up to 5 km long, up to 1 km wide and up to 10 m thick, and ilmenite contents vary between 150 and 300 kg/m³ (Belous 1998).

Alluvial placers of the northwest USH are monomineralic, and contain altered ilmenite (leucoxene). The modern placers are small as a rule and contain fresh ilmenite. Ore bodies in alluvial placers are usually meandering and ribbon-like, and coincide with river valleys. Ore minerals are concentrated in the lower parts of placers. Both residual and alluvial placer deposits are represented by the Irsha group deposits, located in the northwestern part of the USH and worked by the Irsha Integrated Mining and Dressing Combine.

Littoral placers are developed in the central and southern regions of Ukraine and represent ancient and modern sandy beaches and shallow shelves. The former are buried, and the latter exposed. The ancient placers coincide with sandy-clayey sediments of the Oligocene-Miocene seas, and the modern ones formed by long-range drift on spits of land along comparatively short parts of northern coastlines

bearing gabbro-anorthosite massifs in the northwestern part of the USH. The average thickness of the weathering crust is between 10 and 15 m, but within linear tectonic zones it reaches 50-70 m and even more in places. The content of ilmenite in the weathering crust depends on its content in the primary crystalline rocks and normally fluctuates from 30 kg/m³ up to 130 kg/m³, reaching 200 kg/m³ in productive horizons of weathered crust. According to Tarasenko (1992), layered ore bodies consist of kaolinite, montmorillonite and hydromica, and contain 6.0-10.8% TiO_2 and 0.5-2.6% P_2O_5 related to

Deposit	Geological setting	Host rocks and their age	Ore body morphology	Ore mineral composition	Simultaneously useful component
Irsha group deposits	Sedimentary cover of the USH	Alluvial and alluvial-deluvium deposits (secondary kaolin and kaolin clay), Mesozoic and Cenozoic	Layered	Ilmenite	Apatite
Malyshev, Volchansk, Tarasov, Zelenoyar	Sedimentary cover of the USH, littoral placers	Sandy-clayey deposits, Neogene	Layered and lens-like	Ilmenite, rutile, leucoxene	Zircon, kyanite, staurolite, xenotime, sillimanite, diamonds
Krasnokut	The Dnieper-Donetz aulacogen, deltaic placers	Sandy-clayey deposits, Neogene	Layered and lens-like	Ilmenite leucoxene	Zircon

of the Azov and Black Seas. The industrial value of the modern placers is rather limited, and the ancient types form the main industrial titanium deposits. The latter are represented by a number of deposits (Table 2), among which only those at Malyshev are being mined by the Verkhnedneprovsk State Mining and Metallurgical Combine. The principal products are concentrates of rutile, ilmenite, zircon, kyanite-sillimanite and staurolite, as well as glass, and moulding and building sands. The kyanite-sillimanite concentrate contains diamonds 0.01-0.03 mm in size and forming as much as 1 carat/t of concentrate.

The peculiarity of the Miocene littoral placer deposits of Ukraine is the presence of a large number of fine-grained diamonds of kimberlite, lamproite, eclogite and impactite types, as well as some of undetermined (pyroclastic komatiite?) origin (Kvasnitsa & Tsymbal 1998). These placers were formed as a result of erosion and redeposition of sands older in age than those hosting diamonds, the native sources of which remain unknown.

Endogenic deposits

There are four areas with endogenic titanium ores in Ukraine (Fig. 1). Three of them (I, II, III) are related to anorthosite massifs and bodies within the large-scale and composite Proterozoic plutons: Korosten, Korsun-Novomirgorod and East Azov. The fourth area, South Donbass, coincides with Devonian volcano-plutonic rocks situated in the border zone between the Azov block and Dnieper-Donets aulacogen.

There are a number of Ti-deposits-prospects related to gabbro-anorthosite within the Korosten and Korsun-Novomirgorod plutons. Three deposits may serve as examples: Stremigorod, Krapivnia and Nosachev.

The Stremigorod Ti-deposit, situated within the Korosten pluton, is related to the monzonite-peridotite complex of the 830 km² Chepovichi gabbro-anorthosite massif. The ore body has an irregular oval shape (2.3 x 1 km in size), elongated in a northwest direction along the Central Korosten fault. In cross section, the ore body is conical with rather steep and sharp contacts with the host rocks. It is traced to a depth of 1200 m and is characterised by a concentric structure: the central part is represented by plagioclase peridotite and melanocratic troctolite, which pass gradually in the peripheral part to leucocratic troctolite, olivine gabbro, gabbro-monzonite and gabbro-pegmatite (Fig. 2). Troctolite is the main rock type in the ore body.

Ores of the Stremigorod deposit are characterised by uniform compositions and distributions of apatite and ilmenite. Concentration of ore minerals increases from the peripheral to the central parts of the ore body (Table 3). The richest apatite-ilmenite ores in the central part of the deposit contain 6.9-8.17 % TiO_2 and 2.8-4.9% P_2O_5, whereas the peripheral part contains 3.36-5.9% and 0.65-1.5`%, and andesinite and monzonite contain no more than 1% TiO_2 and 0.5 % P_2O_5. The weathering crust above the Stremigorod deposit contains between of 8-10% TiO_2 and 2.5% P_2O_5.

Three varieties of apatite-ilmenite ores are distinguished: (1) poor, disseminated in olivine gabbro, (2) medium-rich, disseminated in leucocratic troctolite, (3) medium-rich, disseminated in troctolite and plagioclase peridotite.

Mineral and chemical composition (%)	Host rock gabbro	troctolite	peridotite
Plagioclase (An_{40-55})	60.6-64.4	37.1-46.0	25.5-31.6
Pyroxene (salite of augite series)	0.6-3.2	0.4-1.4	i.g*-0.7
Olivine (hyalosiderite Fa_{35-45}-hortonolite Fa_{50-58})	3.3-8.4	17.5-22.7	20.1-26.4
Ilmenite	4.95-11.4	11.6-12.6	11.6-15.1
Titanomagnetite	0.4-0.85	1.2-1.9	0.55-3.87
Apatite	1.1-3.0	4.5-6.0	8.1-10.1
Amphibole, talc, chlorite	15.1-19.8	17.9-21.7	16.8-21.7
TiO_2	3.36-5.99	6.8-7.2	6.9-8.17
P_2O_5	0.65-1.5	2.8-3.1	3.58-4.5
FeO	8.97-12.3	14.0-16.47	15.0-21.82
Fe_2O_3	2.82-4.35	5.27-5.86	4.56-6.31
MgO	3.38-6.02	7.37-8.17	6.87-9.01

*i.g.: Individual grains

Table 3. Modal and chemical characteristics of apatite-ilmenite ores of the Stremigorod deposit (after Tarasenko 1992).
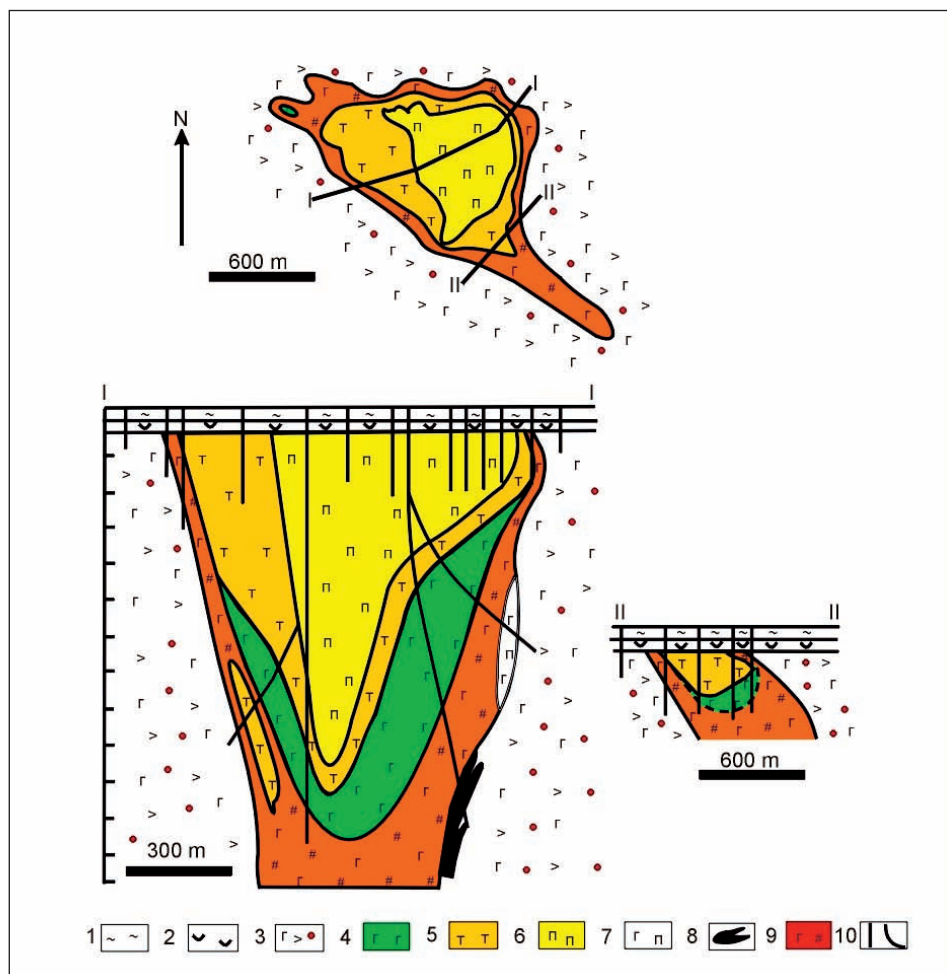
Fig. 2. Schematic geological map and cross sections of the Stremigorod deposit (after Tarasenko 1990). 1 - sedimentary cover rocks, 2 - weathering crust, 3 - K-feldspathised gabbro-anorthosite, 4 - gabbro, 5 - ore-bearing troctolite, 6 - ore-bearing plagioclase peridotite and melanocratic troctolite, 7 - gabbro-pegmatite, 8 - olivine gabbro-pegmatite, 9 - gabbro-monzonite and quartz andesinite, 10 - exploratory borehole.
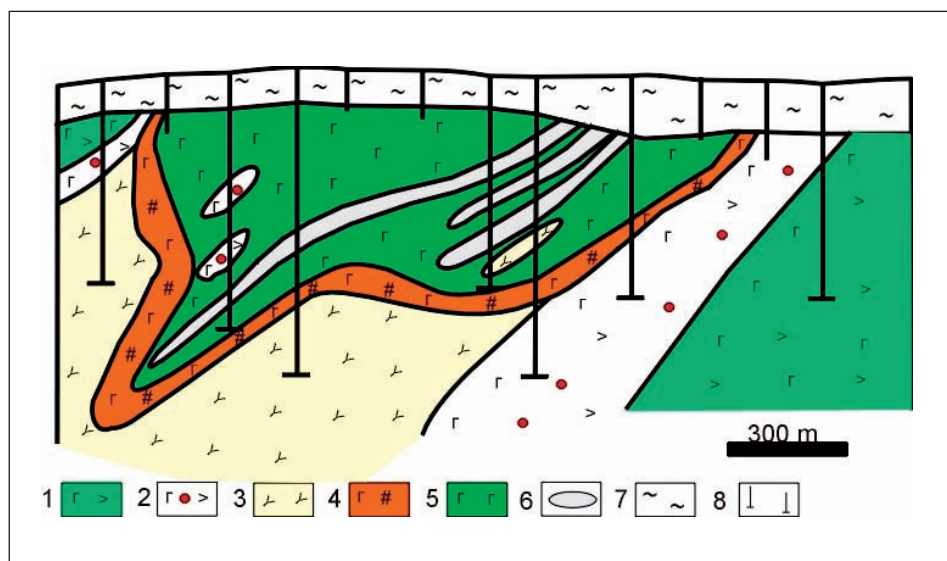
Fig. 3. Schematic geological section of the Nosachev deposit (after Tarasenko 1990). 1 – gabbro-anorthosite, 2 - K-feldspathised gabbro-anorthosite, 3 - quartz andesinite, 4 – monzogabbro-norite, 5 - gabbronorite with disseminated ilmenite ore, 6 - rich, disseminated and massive ilmenite ore, 7 - sedimentary cover rocks, 8 - prospect drillhole.

In the ore-bearing troctolite, 80-84% of all the TiO_2 is concentrated in ilmenite, and 90-93.8% of the P_2O_5 is in apatite. The ores are characterised by a rather high extraction rate of ilmenite and apatite and high contents of TiO_2 and P_2O_5 in the concentrates.

The Krapivnia Ti-deposit is located in the central part of the Volodarsk-Volynsk gabbro-anorthosite massif within the Korosten pluton. The area of the massif is 1250 km^2.

The ore body has an oval shape; it is about 350 m thick, and is characterised by a zonal structure. From the peripheral part of the ore body to the centre, the host rocks change from gabbro-anorthosite to andesinite, leucotroctolite and gabbro, ore-bearing olivine gabbro, ore-bearing olivine pyroxenite and plagioclase peridotite.

The ore minerals (ilmenite, magnetite, titanomagnetite and ulvospinel) are concentrated in the interstices between the silicates, forming a sideronitic texture. Of the whole-rock content of TiO_2 in the ore, 28% is concentrated in ilmenite and 60% in titanomagnetite. The latter also contains 0.41% V_2O_5. Mineral and chemical compositions are given in Table 4.

The Nosachev Ti-deposit, situated within the Korsun-Novomirgorod pluton, is located in the southern part of the Smela gabbro-anorthosite massif. The deposit is lens-shaped, with an overall length of 1.5 km, and a conical section.

Ore-bearing gabbronorite is bordered by gabbronorite, monzo-gabbronorite, quartz andesinite and intensely K-feldspathised gabbro-anorthosite (Fig. 3). The zoned structure of the deposit is emphasized by variations in the chemical composition of the host rocks, including their TiO_2 contents: quartz andesinite (no more than 1% TiO_2), gabbronorite (between 3.68-4.58%), ore-bearing gabbronorite (more than 7%).

In gabbronorite, forming the outer part of deposit, plagioclase is the dominant mineral (50-60%). It is granulated and shows de-anorthisation and a very fine segregation of K-feldspar. The content of disseminated ilmenite reaches up to 7-10%.

Ore-bearing gabbronorite is characterised by medium-rich disseminated and massive ores (Table 5).

The Kalchik Ti-prospect, situated in the southern part of the East-Azov pluton, is typically made up of syenites, including one hosting deposits of REE-Zr and Ta-Nb ores. Anorthosite and rapakivi granite are less characteristic, though the East-Azov pluton is synchronous with the Korosten and Korsun-Novomirgorod plutons. According to Tarasenko et al. (1989), the Kalchik Ti-ores are enveloped by steeply-dipping bodies of gabbroid rock, varying from a few dozens of metres up to hundreds of metres in thickness, and from a few hundred metres up to 4 km and more in length. Ore-bearing gabbroid is represented by layered lenses and schlieren which have indistinct contacts with the host rocks (composite layering of gabbro-pyroxenite, wehrlite, andesinite, gabbrodiorite and diorite). Ore bodies from a few up to tens of metres in thickness and up to a few hundred metres in length are characterised by low and moderate amounts of ilmenite (6-10%), titanomagnetite (5-7%) and apatite (5-7%). Richer mineralizations, but limited in thickness, occur within wehrlite and pyroxenite with titanomagnetite: ilmenite ratios of 1:2 and 1:1, respectively. In spite of the low grade of the ores, economic products of apatite and ilmenite have been obtained by concentration.

The Late Palaeozoic Ti-bearing volcano-plutonic assemblage

The setting of this assemblage is determined by the Volnovakha fault zone in the eastern part of Ukraine, on the boundary of the Azov block and the Variscides of the Donets basin. The latter is the southeastern part of the Dnieper-Donets aulacogen, separating the USH from the Voronezh massif in Russia.

The Late Palaeozoic volcano-plutonic assemblage is developed between the North- and South-Volnovakha faults, as a strip 60 km long and 5 km wide, and the rocks of the assemblage are regarded as a potential Ti-bearing formation (Nechaev 1970).

The assemblage includes three successive complexes (Buturlinov at al. 1972):

Mineral and chemical composition (%)	gabbro	Host rocks peridotite	pyroxenite
Plagioclase An_{45-47} and An_{50-55}	22	6	i.g.
Pyroxene (titanaugite)	29	40.5	45.0
Olivine	14	21	19
Ilmenite	4	6	3.5
Titanomagnetite	17	21.5	25.5
Apatite	4	5	7
Amphibole, talc, chlorite	0.5	i.g.	i.g.
TiO_2	5.84	6.69	7.84
P_2O_5	2.58	2.79	3.17
FeO	20.55	24.66	29.34
Fe_2O_3	4.03	5.44	5.44
MgO	5.96	7.01	7.82

Table 4. Mineral and chemical characteristics of apatite-ilmenite-titanomagnetite ores of the Krapivnia deposit (after Tarasenko 1992).

Mineral and chemical composition	Disseminated	Type of ores Rich disseminated	Massive *
Plagioclase	30-40	25-30	5-10
Pyroxene (ortho- and clino)	20-30	20-30	10-15
Olivine	10-20	i.g.	i.g.
Ilmenite	15-20	30-50	70-80
Titanomagnetite	-	-	-
Apatite	1-3	1-3	i.g.
Amphibole, talc, chlorite	4-7	3-5	5-7
TiO_2	7.0	16.08	34.2
P_2O_5	≤0.26	≤0.25	0.06
FeO	14.89	17.77	29.32
Fe_2O_3	0.9	1.43	1.94
MgO	4.45	3.41	3.95

* Ores of massive type contain 0.24-0.28 V_2O_5 and 0.08-0.12% Cr_2O_3.
  Ores are high-technological: TiO_2 extraction in concentrates reaches 84%.

Table 5. Mineral and chemical characteristics of ilmenite ores of the Nosachev Ti-deposit (after Tarasenko 1992).

1. Intrusions of alkaline-ultrabasic character (390-380 ± 20 Ma),
2. Extrusions of basaltoid (365-335 Ma),
3. Intrusions of nepheline syenite (330 Ma).

The first and third complexes are developed to the east of the Volnovakha fault zone, where they intrude the Precambrian basement and form the Pokrovo-Kireyevo massif, whereas the second is distributed throughout the Volnovakha zone. Complexes 1 and 2 are Ti-bearing, but richer concentrations of TiO_2 are associated with to the alkaline massifs.

Rock	Oxydes (%)								
	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
Picrite-basalt (30)*	41.85	5.18	9.61	7.00	7.09	11.42	10.84	1.72	0.69
Basalt (197)	43.85	5.27	13.18	9.04	5.09	6.40	7.71	2.70	1.99
Andesite-basalt (74)	48.60	4.45	14.63	9.43	3.46	4.50	5.13	3.48	2.21
Trachybasalt (10)	49.79	3.85	15.89	9.10	2.97	3.66	4.08	3.23	4.32
Ore-bearing gabbro-norite of the USH anorthosite massif (after Tarasenko 1987)	49.22	4.60	15.44	1.78	11.67	4.33	7.37	2.70	1.58

* in brackets- number of analyses

Table 6. Chemical composition of basaltoid rocks of the Volnovakha fault zone (after Buturlinov et. al. 1972).

The Pokrovo-Kiryevo massif is oval in form and covers an area of 10 km². It consists of plagioclase pyroxenite and leucocratic gabbro. In pyroxenite along the massif border zone, bodies of ore-bearing pyroxenites, 400-600 m long, contain on average 9.05% TiO_2, 0.049% V and 0.15% P. Ti-bearing minerals are titanomagnetite and ilmenite, as well as titanaugite. Titanomagnetite is the main ore mineral.

The potentially Ti-bearing Devonian basaltoid complex (Table 6) forms a cover, with a maximum thickness of 500 m. Ti-bearing minerals in the main basaltoid rocks are titanaugite, titanomagnetite and ilmenite, with titanite and leucoxene in hydrothermally altered rocks.

There are no known economic Ti-ores related to the basaltoid rocks. However, it may be seen (Table 6) that ore-bearing gabbronorite from the anorthosite massifs is similar to andesite-basalt in Si, Ti, Al, Mg and total Fe contents, as well as to basalt in Ca, Na and K contents.

Conclusions

Ukraine possesses many deposits of titanium ores, among which those in production at present are placer and residual deposits related to weathering of Ti-bearing anorthosite massifs. Placer deposits are the main industrial type of Ti-ores, and contain considerable reserves.

The greatest resources of Ti-ores are related to the Proterozoic anorthosite massifs in the USH. The Devonian Ti-bearing volcano-plutonic complex is interesting from the standpoint of the origin of the geochemical peculiarities of the Proterozoic anorthosite suite.

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Links between the Proterozoic anorthosite-rapakivi granite plutons and ore-forming events in the Ukrainian Shield (ores of titanium, uranium, rare metals and gold)

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Hydrothermal and metasomatic processes have played a major role in the evolution of ore deposits associated with the Proterozoic Korsun-Novomirgorod and Korosten anorthosite-rapakivi granite plutons on the Ukrainian Shield. Magmatic anorthosite-related Fe-Ti-P ores are strongly affected and the ore elements redistributed and enriched by metasomatic processes associated with rapakivi granite intrusions. In addition, hydrothermal systems associated with the rapakivi granites as well as other late-stage acidic intrusions, have formed a large variety of metasomatic/hydrothermal ore deposits; some of these contain considerable resources of U-Be, REE-Zr, Li and accompanying rare metals including gold.

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Introduction

Two large Proterozoic anorthosite-rapakivi granite plutons situated within the Ukrainian Shield (USH) belong to the Palaeoproterozoic Kirovograd orogen-/Central Ukrainian fold belt. Extensive reserves of Ti ores and all economic occurrences of gemmiferous (morion, topaz, beryl) pegmatite are located within these plutons. Numerous deposits and prospects in adjoining plutons contain major reserves of U and Be ores, considerable resources of REE-Zr, Li and accompanying rare metal ores including gold. An analysis of the USH tectonic and metallogenic development, together with spatial and age relationships with the anorthosite-rapakivi granite suite, reveals the leading role of the plutons as an energy/temperature factor for economic ore formation.

Geological setting of anorthosite-rapakivi granite plutons of the USH

The Korsun-Novomirgorod (KNP) and Korosten (KP) plutons, belonging to the anorthosite-rapakivi granite suite, are situated in the central and northwestern parts of the USH, respectively. The KNP, about 5000 km², coincides with the central part of the Palaeoproterozoic Kirovograd orogen (KO). The latter separates the Archaean Podol domain (foreland) to the west from the Dnieper domain (hinterland) to the east. The KP, about 10000 km², is located at the intersection of the

KO and the thrust belt of the Osnitsk-Mikashevichi orogen (Fig. 1).

The Podol domain comprises an enderbite-gneiss complex hosting REE-Zr-Th mineralization, and the Dnieper domain is made up of a granite-greenstone complex with ores of Fe, Au, Ni, Cu and Mo. Borders between the Archaean domains and the KO are represented by thrust zones in the foreland and a system of nappes in the hinterland. The interior part of the KO comprises calc-alkaline associations of island arc affinity. Thrust zones in the foreland, hinterland and corresponding areas of the KO contain BIF, ophiolitic mafic-ultramafic rocks and flysch assemblages forming complexes of intra-continental rift, oceanic crust and passive continental margins. All of these geodynamic settings determine the Palaeoproterozoic metallogeny preceding the emplacement of the anorthosite-rapakivi granite plutons.

The oldest Palaeoproterozoic sedimentary sequence in the eastern part of the KO is represented by coarse Au-U-bearing terrigenous rocks of the Krivoy Rog supergroup. Isotopic ages of U-mineralization fall in the range 2.6-2.45 Ga (Belevtsev 1995) or 2.68-2.31 Ga (unpublished data). The Archaean ages are those of detrital radioactive minerals (erosion products), and the Palaeoproterozoic ones correspond to sedimentary-epigenetic U-mine-

ralization as well as to the first phase of Palaeoproterozoic granitoids in the range 2.45-2.31 Ga (Shcherbak & Ponomarenko 2000, Stepaniuk 2000).

Formation of the KO occurred about 2 Ga ago as a result of collision of an island arc with the Podol domain. This explains why the Palaeoproterozoic granitization is most abundant in the northeastern part of the domain and adjacent to orogenic areas. Ages of monazite from oresegregations within the Podol domain, dated as 3.32, 2.52, 2.29 and about 2 Ga, reflect the inheritance of REE-mineralization from Palaeoarchaean to Proterozoic times. Widespread within the KO, U-ore mineralizations, located in the Palaeoproterozoic metamorphic rocks and represented by metasomatic U-K (K-feldspar) and hydro-

thermal fissure-vein types, have an age of c. 2 Ga (2.19-1.93). The U-K ore type contains relatively high concentrations of REE- and Th-bearing minerals as well as molybdenite (Belevtsev 1995). It is estimated that the ore-regenerative potentialities of K-granite within the KO is mainly due to high U background contents of the granite and to the existence of initial sedimentary-diagenetic U-mineralizations in the rocks which were involved in the formation of the anatectic/palingenetic granites. Besides, related in time and space to these granites are subeconomic spodumene-bearing pegmatite and microcline-apatite- tourmaline- biotite- cordierite metasomatites enriched in Li, Rb, Cs and Be, as well as magnesian skarn and diopside-plagioclase tactite. The latterre presents a metapelite-marl assemblage, initially enriched in W and Au (Kobzar 1981). The spatial connection between Li, B and W-bearing rocks allows comparison with deposits in the USA, where considerable resources of these elements are present in the Neogene deposits of the Cordilleras (California, Nevada), as well as in brines and salt lakes (Silver Peak, Searls and others) (Kozlovskiy 1989). A similar precursor was possible within the Palaeoproterozoic epicontinental depressions in the central part of the USH.

The deep border between both Archaean domains, in accordance with seismic sounding (Chekunov 1992), is determined by the Transregional fault corresponding to the axial zone of the KO (Fig. 1). The zone is characterised by: (1) a decrease in depth to the Moho (between 45 and 35 km only) and a sharp reduction of the basaltic layer (up to 2 km); (2) a decrease in rock density, including those rocks identified as belonging to the upper mantle; (3) the most extensive development within the USH of anatectic K-granitoids and spatially related anorthosite-rapakivi granite plutons.

Fig.1. Tectonic divisions of the Ukrainian Shield, location of anorthosite - rapakivi granite plutons and principal deposits and prospects (excluding numerous occurrences/manifestations of ore mineralisation). 1-3 – Kirovograd orogen: 1-volcanic belt, 2-backarc basin and shelf of hinterland, 3-frontal overthrust belt and parautochthon; 4-5 – plutons (A-Korosten, B-Korsun-Novomirgorod): 4-rapakivi-granite, 5-anorthosite; 6-9 – Faults: 6-Transregional with anatectic granitoids (1); 7-Main overthrust zones of orogen: (2) Odessa-Zhitomir, (3) West Ingulets, (4) Kirovograd, (5) Zvenigorodka-Annovka, (6) Jadlov-Traktemirov, (7) Pervomaysk-Brusilov, 8 – Overthrust zones of backarc Osnitsk-Mikashevichi orogen: (9) Teterev, (10) Perga; 9 – Shear Zones: (11) Krivoy Rog-Kremenchug megashear, (12) Talnov, (13) Central; 10-19 – Deposits/prospects: 10-U hydrothermal vein type, 11-K-U (REE±Mo) type, 12-Na-U type, 13-Ti (P,V,Sc), 14-REE-Zr, 15-Rare metal (Li,Be et all), 16-Mo, 17-W (skarn type), 18-Au, 19-graphite. The metallogenic specialization of the Archaean Podol and Dnieper domains is shown in brackets.

Origin of the anorthosite-rapakivi granite suite

According to petrological studies (Tarasenko 1987) the suite formed as multiphase intrusions with emplacement of subcrustal high-alumina basaltic magmas during phase I, and subsequently during phase II, of crustal granite magmatism. The first phase at 1.8-1.78 Ga (Nechaev &

Boyko 1988, Dovbush & Skobelev 2000) resulted in the intrusion of large anorthosite massifs and related small syenite intrusions/stocks, enriched in Ti, P and REE, Zr, respectively. The lower-crustal rocks of the Archaean-Palaeoproterozoic 'basement' may have caused contamination of the basaltic magma. The second phase, at about 1.77 Ga, is represented by the rapakivi granite unit is followed by the principal Ti ore-bearing complexes at about 1.76 Ga. The granite-porphyry and rhyolite unit, at 1745-1737 Ma (Dovbush & Skobelev 2000), is followed by a veined granite 1685±30 Ma which cuts across the rapakivi granite (Nechaev & Boyko 1988). As has been noted by Dovbush & Skobelev (2000), K-Ar data between 1.8 and 1.7 Ga on biotite and amphibole from rocks of the suite correspond to the same range of ages that have been obtained by U-Pb and Sm-Nd methods. In accordance with Pastukhov & Khvorova (1994), the suite forms a single magmatic assemblage, which is the plutonic equivalent of the basalt-andesite-liparite series, and from such a staindpoint the suite is rather similar to the Devonian basaltoid complex of the Volnovakha fault zone (Gursky et al. 2001).

The value of ε_{Nd} for the Ti ore-bearing complexes is negative (-0.8), and similar to the value calculated for anorthosithe belonging to the early unit (Dovbush & Skobelev 2000). The ε_{Nd} for rapakivi granite is below –1.8, and both the mafic and the felsic units therefore could not be directly derived from depleted mantle. They are most likely derivatives of partial melting of the lower crust and/or the upper mantle, enriched by oceanic crustal material due to a previous subduction. Melting of high-alumina basalt could produce the initial magmas for mafic rocks belonging to the plutons. The rapakivi granites cannot be derived from the same parental magma, as their isotopic signatures differ from those of the mafic rocks. It seems likely that anatexis of the upper crustal rocks stimulated rapakivi granite formation and, potentially, the ore-bearing features of syenite spatially and temporally related to anorthosite (gabbro-syenite intrusive complex after Nechaev & Krivdik 1989). Isotopic data supported by geophysical deep sounding (Chekunov 1992) turn our attention to the probability that the basaltic layer was a principal factor for the formation of anorthosite as well as for lateral tectonic transference within the KO.

The spatial and age relationships between plutons and ore formation

As noted above, the potential ore-bearing mineral complexes were formed before emplacement of the plutons. Mineral-chemical ore types of economic rank which were generated by pluton emplacement indeed occur in those geological settings where their corresponding precursors were already present. The best example of this kind is displayed by REE-Zr and Ti ores related to the formation of the anorthosite-rapakivi granite suite.

Rare earth element-zirconium ores

A deposit of this type of ore is located in the northwestern flank of the KP (Fig. 1). The ore-bearing syenite is characterised by a high background of Zr (0.2-0.3%), and separate ore bodies contain more than 6% ZrO_2. The zircon crystals have complex structures, typically including cores with an age of 1.79 Ga, surrounded by an overgrowth reflecting an overprint of the ore-forming episode at 1.73-1.72 Ga (Nechaev et al. 1985). REE minerals are represented by britholite, orthite and bastnaesite (accompanied by Y-fluorite) and there are two obvious generations of REE-mineralizations. Similar ore prospects are located in the southeastern corner of the KNP (Fig. 1) and in the eastern USH, where the great Azov REE-Zr deposit has been discovered recently.

Titanium ores

These ores are located directly within the primary Ti-bearing anorthosite massifs (1.8 Ga in age) and Ti-deposits are coincident with the Central fault (Fig.1). The principal Ti ore-bearing complexes (1.76 Ga in age) are metasomatic in origin. They consist of monzonite-norite, monzonite-peridotite and oligoclasite-andesinite (Tarasenko 1987, 1990, 1992), and were formed as a result of the contact influence of rapakivi granite on anorthosite during phase II (1.77 Ga). These complexes are located within anorthosite massifs affected by extensive cataclasis and K- feldspathization, and have a zoned metasomatic structure in which ore- and rock-forming components have been redistributed. Ti, Fe and P were leached from the internal

portions of metasomatic columns and deposited at the front of the columns, resulting in ore-mineralization (i.e. not initially magmatic impregnation). The oligoclasite-andesinite complex essentially represents chemical and mineral transformation of initial anorthosite, and zones of basification related to that complex include ilmenite ore containing up to 20-35% TiO_2.

According to Tarasenko (1987), zonation of Ti-bearing complexes is determined by their temperature of formation that depends on the distance from the rapakivi granite contact. The monzonite-norite complex represents the highest temperature (830°C) and is located in the peripheral part of the anorthosite massifs near the contact with rapakivi granite, whereas monzonite-peridotite formed at some distance from the rapakivi granite (750-700°C). The oligoclasite-andesinite complex, remote from the rapakivi granite, was formed at 500-450°C, and is characterized by a high degree of oxidation and leucoxenization of ilmenite compared to the Ti-ores of the monzonite-norite and monzonite-peridotite complexes. The relationships between epigenetic Ti-ores on the one hand, and anorthosite and rapakivi granite on the other hand, imply that anorthosite provides the source of the Ti and the rapakivi granite intrusion appears to be responsible for the metasomatism which caused the redistribution and concentration of titanium.

Epigenetic Ti-ores within anorthosite massifs are only one phase in a sequence of ore formation episodes which spread outside the plutons. Emplacement of anorthosite and extensive K-feldspathization, leading to de-anorthization of plagioclase, caused Ca- and Na-metasomatism, the effects of which extended around and far from the plutons.

Rare metal ores

Deposits of economic rare metal ores occur in the southwestern part of the KNP (Fig. 1). They are represented by extensively brecciated microcline-albite-spodumene-bearing pegmatite with superimposed petalite mineralization, and are considered to be similar to replacement dykes (Nechaev et al. 1991). The complexity of these deposits is caused by extensive metasomatic transformations (c. 1.9 Ga), not only in the initial pegmatite, but also in the host gneissic metasomatite enriched in rare alkalies. The main ore stage at c. 1.8 Ga was synchronous with the KNP emplacement, which was the cause of the high-temperature Li metasomatism: the quartz-albite-petalite assemblage, typical of ore deposits, was formed by solutions at T≤680°C (Voznyak et al. 2000).

The Perga rare metal deposit, located northwest of the KP (Fig.1), contains economic Be-mineralization (genthelvite, phenacite). Ore formation in the period 1.74-1.62 Ga (Shcherbak 1978) is related to K-Na metasomatism with Li-F specialization (Metalidi & Nechaev 1983, Buchinskaya & Nechaev 1990).

Uranium ores

Deposits of metasomatic U-Na (albitite) type occur in K-granites which adjoin the KNP to the south (Fig. 1), as well as in ferruginous rocks of the Krivoy Rog Supergroup remote from the pluton to the east. These ores have an age between 1.8 and 1.7 Ga, and two stages of albitization are separated by a cataclasis (Belevtsev 1995). High-temperature (450-480°C) garnet-diopside albitite, which is only developed near the contact with the KNP (Belevtsev 1995), is actually, in our opinion, calcareous skarn that has suffered Na-metasomatism. Their formation temperature is close to the oligoclasite-andesinite Ti ore-bearing complex (450-500°C), and the albitite halo of the KNP appears to be a continuation of Na-metasomatism within the pluton itself, whereas most of the U-Na ore type formed at 250-300°C.

A geochemical feature that links the Ti- and U-ores is the presence of accompanying elements such as P, Zr, V and Sc. Primary ilmenite in anorthosite contains an average of 750 ppm V and 56.5 ppm Sc, whereas ilmenite in Ti-ores contains averages of 1550 ppm and 74 ppm, respectively (Borisenko et al. 1980). The early mineral paragenesis of U-bearing albitite includes apatite, malacon and U-titanates (brannerite and davidite). Andradite from garnet-diopside albitite contains up to 3% TiO_2 as well as about 1% V, and this albitite also contains tortweitite with > 40% Sc_2O_3 and 7.6% ZrO_2. The complex Sc-V-U ore of the Zheltaya Rechka deposit is related to Na-metasomatism of ferruginous rocks of the Krivoy Rog Supergroup,

where aegirine-acmite contains 0.08-0.1% Sc and 2-9% V (Belevtsev 1995). It may be presumed that the Ti, V, P, Zr and Sc sources belonging to the U-Na ore type were hydrothermal solutions related to the pluton paragenesis.

Gold ores

The main deposits and prospects of Au ores known at present are situated in the southern part of the KNP (Fig. 1). All are associated with U and rare metals as well as with tourmaline-bearing fields, and commonly contain scheelite.

Epigenetic Au ores formed after extensive deformation, including mylonitization of the country rocks (amphibolite, gneiss, migmatite, granite, skarn, aplite and pegmatite). Au-bearing mineralizations are located within linear zones of brittle deformation. The interior parts of these zones are affected by oligoclase-quartz-biotite metasomatism and contain ore mineral assemblages with native gold and tellurides (far from the KNP) and maldonite without tellurides (close to the pluton).

Alteration of the host rocks by hydrothermal ore-bearing solutions is displayed by minor albitization of plagioclase and formation of K-feldspar, carbonates and mixed-layer silicates of illite-smectite type (Mudrovskaya 2000). The upper temperature for the formation of the Au-bearing assemblage corresponds to the transformation of hexapyrrhotine into clinopyrrhotine (273°C) and the lower temperature limit is fixed by the formation of mixed-layer silicates (<178°C). The thermal parameters of Au-ore formation are similar to those of the post-albitite assemblages of U-Na ore deposits. These assemblages formed at temperatures lower than 300°C under the influence of hydrothermal solutions, characterised by high activities of K and CO_2 compared to U-bearing albitite (Belevtsev 1995). The Au ore-forming processes therefore appear to be a continuation of the U-Na ore-forming event. On the other hand, the geochemical antagonism between Au and U within ore-bearing areas, ore fields and even deposits can be explained by the transfer of Au as hydrosulphide complexes. The formation of Au-ores after U-ores agrees with the occurrence of Au-bearing assemblages accompanied by adularia-chlorite metasomatism which cuts across U-bearing albi-

tite (Prokhorov et al. 1994). However, according to Yatsenko & Gurskiy (1998), Au-bearing mineral assemblages (550-250°C) developed after tourmaline- and scheelite-quartz (420-330°C) in the KNP region. This temperature inversion probably reflects the influence of the pluton. Au dispersed in Fe and As sulphides (550-380°C) subsequently changed to an assemblage of native gold + chalcopyrite (450-370°C) and then to native gold + bismuth. (c. 270°C). This suggests that Au mineralization + tellurides located far from the KNP appear to be a continuation of the higher temperature Au-bearing assemblages located close to the pluton.

Waning of magmatic activity in the KNP is fixed by granite veins which cut across rapakivi granite at 1685±35 Ma, and also by nearly synchronous retrograde metamorphism (1670±30 Ma) of gneiss surrounding the pluton. Epigenetic quartz veinlets including Au-Bi-As mineralization occur in the granite veins, and the K-Ar age of biotite from a Au-bearing oligoclase-quartz metasomatite distant from the pluton is 1615±30 Ma. The timing of events related to economic Au-mineralization indicates that the KNP was a significant thermal factor for ore formation.

Untraditional occurrences of Au related to REE-Zr-bearing syenite and Nb-Sn-bearing granite also occur in the northwestern part of the KP within the Perga rare metal ore field (Nechaev et al. 1983, 1986), and precious metal mineralization is considered to belong to a rare metal-polymetallic evolution series (Nechaev 1990) . The latter mineralization could be a result of regressive hydrothermal processes caused by the influence of the KP (1.8-1.7 Ga) because syenite, granite and Be-bearing metasomatite are synchronous and spatially related to the pluton.

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Approaches and processes

Do Fe-Ti oxide magmas exist?
Geology: Yes; Experiments: No!

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Field evidence strongly suggests that many Fe- and Fe-Ti-oxide bodies were emplaced as liquids. The evidence is strongest for the deposits at El Laco, Chile, where magnetite bodies appear to have been extruded as lavas, and for which the experiments of Weidner (1982) in the system Fe-C-O provide convincing proof of oxide liquids at geologically reasonable temperatures (<1000°C). Thus, it was very exciting when we discovered that the Fe-Ti-oxide bodies in the Laramie Anorthosite Complex contain graphite. Very importantly, however, Weidner's experiments lacked TiO_2 - and the El Laco iron-oxide lavas are nearly titanium-free.

Field evidence of a magmatic origin for Ti-bearing oxide bodies (ilmenite and magnetite-ilmenite ores) still seems strong: many are dike-like, crosscutting the country rock with sharp contacts. Many are associated with anorthosite complexes. Isotopic studies show that many are closely akin to the host anorthosites. Many of these bodies ('nelsonites') contain abundant apatite, and it has been suggested that apatite serves as a flux to stabilize Fe-Ti-O-(apatite) magmas, possibly as melts immiscible with a complementary silicate liquid.

Unfortunately, despite much effort there is no experimental evidence to support the concept of apatite-fluxed melts. Philpotts (1967) reported ilmenite-apatite melts at ~1400°C, but this is hardly compelling. The temperature is unrealistically high for crustal conditions, and in any case fluorapatite and ilmenite separately melt near that temperature, so there is little evidence for a eutectic or cotectic relationship between them. Kolker (1982) pointed out that while nelsonites broadly contain approximately 1/3 apatite and 2/3 oxide, the proportions in fact vary widely, further casting doubt on the existence of a eutectic relationship.

For many years my students and I have tried - unsuccessfully - to generate Fe-Ti oxide liquids in the laboratory at geologically reasonable temperatures. We have tried apatite and iron phosphate and graphite - separately and in combination - as well as chlorides as possible fluxes. Using graphite alone we have made small amounts of Fe-rich liquid - but importantly that melt is nearly Ti-free; most of the titanium remains in the residual crystals. Using graphite and apatite we again made small proportions of Fe-rich, Ti-poor melt - but that melt dissolved more strontium than phosphorus from the natural apatite used in the experiments! Despite the common association of apatite and oxides in nature, there is no experimental support for the existence of apatite-oxide melts.

The experiments of Epler on an oxide-rich troctolite pegmatite (MS thesis, Stony Brook, see also Lindsley et al. 1988) suggest a possible explanation. Although he failed to produce an immiscible oxide *melt*, he showed conclusively that phosphorus and iron *mutually enhance their solubilities* in the silicate melts coexisting with crystalline Fe-Ti oxides. For example, at 5 kbar, 1100°C he produced a silicate liquid with 26.2 weight% iron as FeO, 4.2% TiO_2, and 3.8% P_2O_5. At 1150°C the values go up to 32.7, 8.56, and 4.2%, respectively. Thus, the close association of apatite and oxides probably reflects coprecipitation from a silicate parent: once the silicate melt becomes saturated with one of these phases, the other is likely to coprecipitate as well. Experiments suggest that the oxide bodies crystallize from Fe-Ti-P-rich silicate liquids and are emplaced eit-

her as crystal mushes or in the solid state.

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Some results on the role of P, T and fO$_2$ on ilmenite composition

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Using experimental and geochemical data acquired on fine-grained samples, representative of liquid compositions, Vander Auwera et al. (1998) have defined for the Rogaland AMC suite, a complete liquid line of descent ranging from primitive jotunites (hypersthene-bearing monzodiorites) to evolved jotunites and then to charnockites. Primitive jotunites represent the least differentiated compositions (high MgO, low K$_2$O) which in the Rogaland Anorthosite Province correspond to the chilled margins of leuconoritic (Hidra: Demaiffe & Hertogen 1981) and layered (Bjerkreim-Sokndal: Wilson et al. 1996) bodies. Evolved jotunites represent more differentiated compositions and occur in the dyke system which crosscuts the Rogaland Anorthosite Province. Modelling of the liquid line of descent supports the hypothesis that extensive fractionation of primitive jotunites can produce quartz mangerites with REE concentrations in the range of jotunites, strong depletions in U, Th, Sr, Ti and P, and smaller to no relative depletions in Hf and Zr.

Jotunites characterised by very high concentrations of FeO, TiO$_2$ and P$_2$O$_5$ (FTP rocks) have also been observed in Rogaland. Experimental and petrographic data indicate that these rocks represent accumulations of a dense oxide-apatite-pigeonite assemblage into coexisting multisaturated jotunitic to mangeritic liquids.

Experimental data performed in anhydrous conditions from 1 atm (fO$_2$ at NNO, FMQ-1 and MW-2.5) up to 13 kb (FMQ-2, FMQ-4) on the so-called Tjörn (TJ) primitive jotunite (Vander Auwera & Longhi 1994) are used to assess the possible role of pressure, temperature and fO$_2$ on ilmenite composition (Al, Mg, Cr). Indeed, in Rogaland, it has been shown that the parent magmas of the Ana-Sira massif-type anorthosite and the layered mafic intrusion of Bjerkreim-Sokndal, which both contain ilmenite deposits (Duchesne 1999), are generally similar to this primitive jotunite (Duchesne & Hertogen 1988, Vander Auwera & Longhi 1994, Vander Auwera et al. 1998, Longhi et al. 1999). These experimental data can thus bring information concerning the behaviour of poisonous elements like Mg and Cr during polybaric crystallization of ilmenite in anorthosite complexes.

In the TJ primitive jotunite, ilmenite is a near-liquidus phase and experiments corresponding to the first appearance of this phase at a given pressure have been selected in order to limit the extent of liquid compositional variability. These experimental data indicate that the most important effects are an increase in D$_{Al2O3}$ (Ilm/liq), a slight increase of D$_{Fe/Mg}$ (Ilm/liq) and a decrease of ilmenite Cr$_2$O$_3$ with increasing pressure. Nevertheless, this latter effect could also result from the
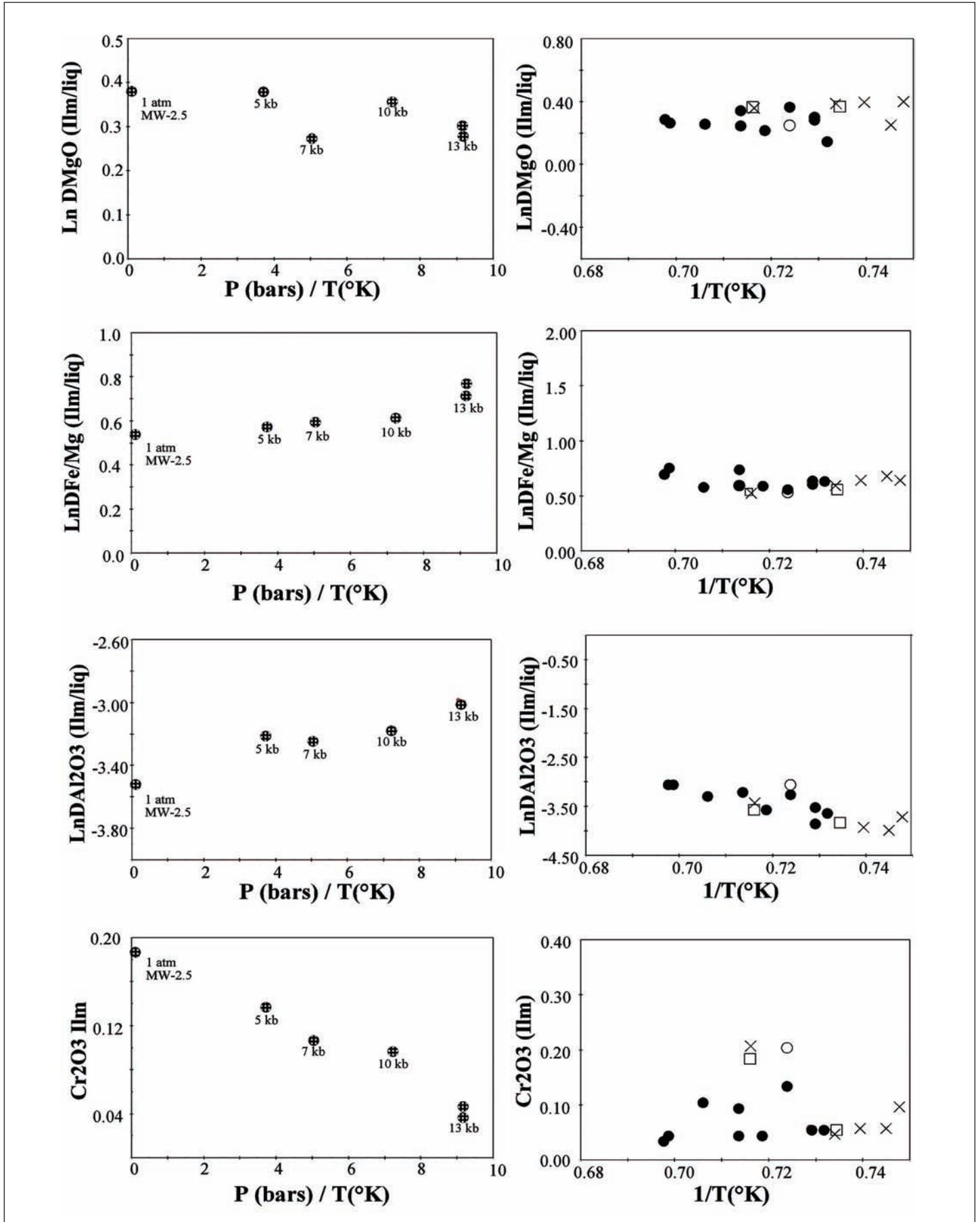
Fig. 1. Partition coefficients of several elements as a function of P/T and T. Open circles: 1 atm, MW; open square: 1 atm, MW-2.5; crosses: 1 atm, FMQ-1; filled circle: 5 to 13 kb.

earlier appearance of opx with increasing pressure, which decreases the Cr content of the liquid. Besides that, there is no significant variation of D_{MgO} with increasing pressure and D_{Al2O3} only slightly increases (Fig. 1). Comparison between observed (Duchesne 1999) and experimental ilmenite compositions show that these ilmenites display similar contents in Fe^{3+} and Mg but experimental ilmenites are usually higher in Al and lower in Mn.

These experimental data thus suggest that the compositional variability observed in ilmenites from different orebodies (Duchesne 1999) probably result from variable parent magma compositions, as these ilmenites may have crystallized at different times along the liquid line of descent. Subsolidus readjustments may also be superimposed on this primary effect (Duchesne 1972, 1999).

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The nelsonite problem: an origin by melt immiscibility

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The origins of ore-bearing apatite rock (nelsonite) from the Proterozoic Suwalki anorthosite massif and Fe-Ti-P-rich hypersthene monzodiorite (jotunite) from the Suwalki and Sejny massifs, NE Poland, have been investigated by means of melt inclusion techniques.

The investigated rock samples come from a deep borehole drilled in the crystalline Precambrian basement in NE Poland, belonging to the East European Craton (Fig.1). An E-W trending shear zone of post-collisional origin controls the occurrence of Meso-Proterozoic intrusions – called the Mazury Complex – of bimodal composition, anorthosite-norite and rapakivi-like granite. The Suwalki and Sejny intrusions are classified as 'massif-type anorthosites', belonging to the AMCG suite of Proterozoic plutonism (Wiszniewska et al. in press). The ore-bearing apatite dykes (nelsonites) were found for the first time in Poland (Lopuchowo boreho-
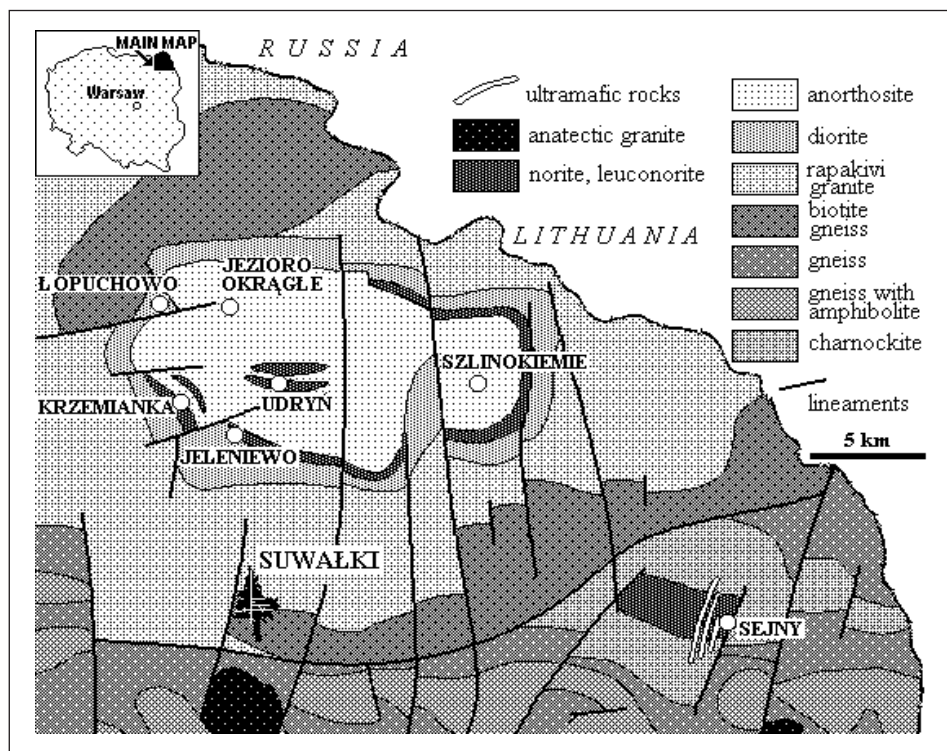
Fig. 1. Geological map of the Suwalki anorthosite intrusion (after Kubicki & Ryka 1982).

le) in the marginal part of the Suwalki anorthosite massif. Jotunite with chilled texture has been found in the Sejny noritic intrusion (Sejny borehole) and also as fine-grained dykes in the Suwalki anorthosite massif (Udryn borehole) (Fig. 1).

Nelsonites consist mainly of fluorapatite (up to 50-70 vol. %), magnetite, ilmenite, pyrrhotite, pyrite and chalcopyrite (Fig. 2). The oxide/sulphide ratio is variable, ranging from 4:1 to 0.7:1. The presence of REE-bearing minerals, like Ce, Nd, La-allanite and rhabdophane (Nd, Ce, La)-$PO_4 \cdot H_2O$, was recognised as rims around apatite grains or as an interstitial product (Wiszniewska 1997).

Jotunite dykes consist of calcic plagioclase, low- and high-Ca pyroxenes, biotite, apatite and Fe-Ti oxide ores as essential components, with local porphyritic structure. The jotunite dykes from the Suwalki massif and chilled zones from the Sejny intrusion have a high content of iron (from 16.5 to 18.8% Fe_2O_3), titanium (from 2.3 to 3. 5% TiO_2) and phosphorus (from 0.6 to 1.2% P_2O_5), a low magnesium number (# mg ca 0.36), and REE distributions moderately differentiated with (La/Yb)-CN c.10 and without a Eu anomaly (Wiszniewska et al. in press). These geochemical characteristics are very similar to those of the Rogaland jotunites (Duchesne 1999, Vander Auwera et al. 1998).

Melt inclusions in nelsonites

Fluid inclusions were found in zoned apatite grains in inner cores, in intermediate zones and in outer rims (Fig. 2). The inclusions in each zone differ in the type of filling and in the homogenisation temperatures: (1) Inner core inclusions are made up of pyroxene, apatite, carbonate, plagioclase, biotite, halite, sylvite and aqueous solution, and homogenise at 890-870°C. (2) The intermediate zones of the apatite grains bear inclusions poorer in pyroxene and plagioclase. They have carbonate and apatite contents similar to the inner core inclusions, but show an increased amount of aqueous solution. Their homogenisation temperatures range from 820 to 770°C. (3) The outer rims contained inclusions rich in solution and in halite, but poorer in carbonates and without silicates. They homogenised at 690-550°C. (4) Gas-liquid inclusions were found in the REE-rich interstitial product. They are filled with aqueous solution, sometimes containing halite and calcite. The homogenisation temperatures ranged from 480 to 230°C.

The homogenisation process in the high-temperature microscope heating stage of inclusions in the core and intermediate zones of the apatite reveals, by increasing temperature, the formation of two melts separated by a meniscus. The qualitative composition of the glassy phases was determined by electron microprobe analysis. The data show that one melt is silicate-rich and the other phosphate-rich. Moreover, small dispersed grains of silicates were also identified and the silicate and phosphate melts were both found to contain carbon (i.e. carbonate component), the carbon content in the phosphate melt being distinctly higher than in the silicate melt. Chlorine was also detected in both melts with an irregular distribution. These results suggest that apatite and the other minerals of nelsonite possibly crystallised from a melt immiscible with a silicate melt. Both melts neither homogenised up to temperature of 1080°C nor even distinctly changed their volume proportions.

Melt inclusions in jotunites

Pyroxenes from coarse-grained jotunites in the central parts of the dykes contain melt inclusions with crystallised fillings consisting of pyroxene, feldspar, magneti-

te, calcite and apatite plus gas bubbles. The crystal aggregate starts to melt to an observable degree at 800–820°C, the last crystals disappear at 980–1040°C, and small droplets of phosphate and carbonate melt form. Complete homogenisation of all phases occurs at 1090–1170°C. Fine-grained pyroxenes from chilled facies contain melt inclusions with fillings being an aggregate of tiny grains of non-recognisable minerals with gas bubbles. These fillings also start to melt at 770–780°C, the last crystals disappearing at 1000–1060°C with the appearance of small droplets of phosphate and carbonate. Complete homogenisation of these droplets and gas bubbles with the silicate melt occurs at 1090–1180°C. Some pyroxenes also contain rare inclusions filled with carbon dioxide, probably in a liquid state.

Conclusions

Observation of the homogenization of melt inclusions point to the formation of the Suwalki nelsonite from the crystallization of a melt rich in phosphate and containing carbonate, which was immiscible with a silicate-rich melt. As inferred from the melt inclusion fillings, the proportion of these two melts was close to a 1:1 ratio. This composition (50% apatite–50% diorite) is quite far within the miscibility gap, ant thus in the two-phase region on a Philpotts`s (1967) diagram. The homogenisation temperature of the fluid inclusions in zoned apatite distinctly indicate that nelsonites formed from a melt which probably gradually and smoothly changed with decreasing temperature to a very saline aqueous solution, then to a moderate-saline medium-temperature aqueous solution forming the latest outer rims of the apatite grains. As for the parent melt of jotunites from Sejny, since its composition plots in the one-phase silicate melt area of the Philpotts diagram, the observed silicate and phosphate melts' immiscibility at high temperature possibly results from a metastable process.
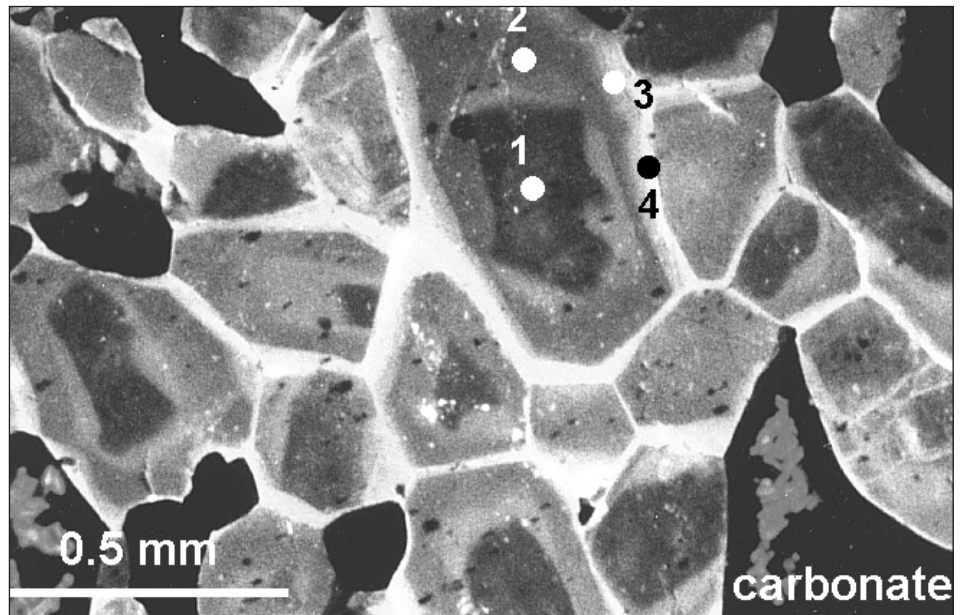
Fig. 2. Cathodoluminescence zoning of apatite from the Suwalki nelsonite; the numbers refer to the fluid inclusion types in apatite in individual growth zones, the circles indicate the typical inclusion positions.

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On ilmenite and rutile ore provinces in Norway, and the relationships between geological process and deposit type

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There are a considerable variety of different types of titanium ore found in Norway (Korneliussen et al. 2000a, 2000b). There is also a close link between geodynamics, at all scales, and the evolution of titanium deposits. At a regional scale, the distribution and character of the Ti-deposits are linked to distinct geological events at the western and south-western margin of the Fennoscandian Shield, ranging in age from Proterozoic to Permian. The magmatic, metamorphic and structural processes involved place distinct fingerprints on the individual Ti-deposits.

From an economic-geological point of view, the mineralogical variations are of overall importance, since mineralogy to a large extent defines the mineability of a Ti deposit. Mineralogy is a function of the geological evolution of the deposit, which is often very complex. In fact, a major task in continued investigations of Ti mineral deposits is to identify those that have ore-mineral qualities that will make it economic to produce titanium mineral concentrates of ilmenite and rutile of sufficient quality, particularly for the Ti-pigment industry.

The dominant Ti-ore category is that of magmatic deposits of ilmenite, titanomagnetite and apatite in various proportions, and deposits variably affected by secondary, metamorphic processes. The major ilmenite province is the Rogaland Anorthosite Province in southern Norway. It was emplaced in a Middle Proterozoic, migmatised, metamorphic terrane at intermediate crustal levels. Titanium deposits are widespread in the province, principally as ilmenite-magnetite-rich cumulates, or as ilmenite-magnetite-rich intrusions. There is a wide variety in chemical characteristics and mineral textures in these deposits.

In some regions, metamorphic processes related to major crustal events have transformed ilmenite-bearing rocks into rutile-bearing rocks. Since rutile is a much more valuable mineral than ilmenite, such transformation processes are of significant economic interest.

In western Norway, including the Sunnfjord province, Proterozoic mafic and felsic igneous rocks were strongly affected by the Caledonian orogeny. Two types of Ti deposits occur in this region: magmatic ilmenite-magnetite deposits associated with Proterozoic mafic intrusions, and rutile-bearing Caledonian eclogitic rocks. The rutile-bearing eclogites are Proterozoic basic rocks that were transformed into eclogite at high pressure during the collision between Baltica and Laurentia c. 400 Ma ago. During this process, ilmenite in the protolith broke down, with the Fe entering garnet and Ti into rutile. Large volumes of ilmenite-bearing mafic rocks were transformed into rutile-bearing eclogitic rocks with the same TiO_2 content. From a mineral resource perspective, the ilmenite deposits within the eclogite regions in W. Norway are of only minor interest, whereas the rutile-bearing eclogites represent a major mineral resource, particularly the Engebøfjell deposit (McLimans et al. 1999).

The third ore category is a variety of the Proterozoic rutile-bearing, scapolitised and albitised rocks in the Bamble - Arendal province (also called the Bamble Sector of the Fennoscandian Shield). These mafic rocks were extensively affected by alkali-metasomatism in the form of scapolitisation and albitisation. In this process, ilmenite in the gabbroic protoliths broke down to rutile during the metasomatic process, while iron was removed from the system by fluids. Thus, metasomatism creates rutile ores.

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Ilmenite–zircon relationships in meta-anorthosites. Example from the Bergen Arc (Caledonides of W Norway) and implications for zircon geochronology

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The Lindås Nappe is situated in the Caledonides of West Norway and composed mainly of a Proterozoic anorthosite–charnockite complex. It was affected by penetrative Sveconorwegian granulite-facies metamorphism, followed by a fluid-driven eclogite- and amphibolite-facies Caledonian overprint. This overprint is spatially restricted along fractures and shear zones (Austrheim & Griffin 1985, Cohen et al. 1988, Boundy et al. 1992, Boundy et al. 1996, Austrheim et al. 1997, Bingen et al. 2001b). In this study, formation of metamorphic zircon was investigated in mafic lithologies. Petrographic observations with optical microscopy, backscattered electron imaging and catho-doluminescence imaging were performed in thin- sections of whole-rocks and polished mounts of zircon separates. SIMS analyses (Cameca IMS1270) were realized on zircon for age determination and estimation of trace-element concentrations (Th and REEs).

In mafic granulites and amphibolites, a luminescent anhedral zircon overgrowth surrounds a magmatic zoned core. The overgrowth gives an average age of 924 ± 58 Ma with a Th/U = 0.52 and the magmatic core gives an age of 952 ± 32 Ma with a Th/U = 1.27. The REE pattern of zircon in a two-pyroxene granulite displays a Ce positive anomaly and enrich-

ment in HREE. The REE pattern of the luminescent overgrowth is parallel to that of the magmatic core, but is characterised by distinctly lower concentrations. In the granulites, a continuous rim of zircon or a discontinuous corona of c. 10 μm rounded to flat zircon crystals is commonly observed in thin-section at the outer margin of ilmenite grains. Baddeleyite (ZrO_2) and srilankite (Ti_2ZrO_6) blebs are reported around ilmenite included in feldspar or pyroxene. Baddeleyite is interpreted as an exsolution product from magmatic ilmenite. Srilankite was probably formed as a reaction product between ilmenite and baddeleyite during granulite-facies metamorphism in silica-deficient subsystems. The zircon corona around ilmenite and the luminescent zircon overgrowth were also formed as a reaction product during granulite-facies metamorphism, where silica was available at grain boundaries. Textures suggest that magmatic ilmenite was a main source of Zr to form metamorphic zircon (Bingen et al. 2001a).

In massive amphibolites, relict ilmenite grains are surrounded by a corona of titanite and a discontinuous corona of micro-zircons. Amphibolite-facies overprint is not associated with any significant growth or dissolution of zircon.

An unsheared eclogite displays a zircon

population with a euhedral oscillatory-zoned overgrowth showing well-terminated edges and tips. The overgrowth yields an age of 424 ± 5 Ma (weighted average of 31 analyses). It is characterised by a Th/U ratio lower than 0.13, a Ce positive anomaly and an enrichment in HREE. The REE pattern of the overgrowth is not significantly different from that of the Proterozoic core, although some of the analyses of the overgrowth are specifically poor in all trace elements. In thin-section, a corona of micro-zircon grains is observed at some distance around rutile, and locally these micro-zircons show a prismatic overgrowth. The specific low-Th zircon growth event is related to eclogite-facies forming reactions, involving breakdown of a two-pyroxene + garnet + plagioclase + ilmenite assemblage to form a garnet + omphacite + rutile assemblage in the presence of a fluid. The low Th content of this zircon probably stems from the coeval precipitation of clinozoisite. This oscillatory zoned zircon records fluid infiltration and coeval Caledonian eclogitization in the crust.

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Evidence from Re-Os for the origin of sulphide concentrations in anorthosites

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Recent Re-Os studies of sulphide concentrations in anorthosite-dominated plutonic suites consistently yield high initial ^{187}Os/^{188}Os ratios in sulphide minerals, indicating a major component of crustal Os. These studies have led to three distinct hypotheses for the origin of the sulphide mineralization, with implications for the genesis of the hosting plutonic rocks: (1) the source of the parental magmas is mafic continental crust; (2) mantle-derived melts achieve sulphide saturation and acquire crustal Os through bulk assimilation of silicate crust; or (3) mantle-derived melts achieve sulphide saturation and acquire crustal Os through selective assimilation of crustal sulphur and metals by volatilization and/or melting of crustal sulphides.

Table 1. Some isotopic characteristics of sulphide-bearing intrusions in anorthosite-related plutonic suites.

Site (~ Age, Ma)	γ_{Os} sulfides	γ_{Os} other[a]	ε_{Nd}	$(^{87}Sr/^{86}Sr)_i$	$\delta^{34}S$ ‰
Suwalki (~1560) [1]	+642 to +962 [1]	+743 to +961 [1]	-6.3 to -1.6 [2]	0.7046 to 0.7055 [3]	~ 0 [4]
Rogaland (~930) [5]	+419 [6]		[b] 0.0 to +4.6 [7]	[b] 0.7033 to 0.7062 [7]	--
[c] Voisey's Bay (~1330) [8]	+898 to +1127 [9]	<0 to +84 [9]	-4.1 to +0.9 [10]	0.7034 to 0.7040 [10]	-4 to +2 [11]
[d] Duluth (~1100) [12]	+500 to +1200 [12]	+3 [12]	-7.1 to -3.4 [12]	--	+10 to +14 [12]
[d] Sally Malay (~1845) [13]	+950 to +1300 [13]	+60 to +479 [13]	-0.2 to +0.7 [13]	--	--

References:
[1] Morgan et al., 2000; [2] Claesson & Ryka, 1999; [3] Wiszniewska et al., 1999b; [4] Wiszniewska & Jedrysek, 1998; [5] Schärer et al., 1996; [6] Schiellerup et al., 2000; [7] Demaiffe et al., 1986; [8] Amelin et al., 1999; [9] Lambert et al., 2000; [10] Amelin et al., 2000; [11] Ripley et al., 1999; [12] Ripley et al., 1998; [13] Sproule et al., 1999.
Notes:
[a] For Suwalki, values are for magnetite, though microscopic sulphide inclusions may be present. For Rogaland and Voisey's Bay, values are for related unmineralized intrusive rocks; negative numbers indicate disturbance of the Re-Os system (Lambert et al. 2000). For Duluth and Salay Malay, values are for weakly mineralized rocks or disseminated sulphide spatially removed from major concentrations. See text for discussion.
[b] Schiellerup et al. (2000) report a much tighter range of ε_{Nd} = 0.22 to 1.99 (n = 30) and $^{87}Sr/^{86}Sr$ = 0.705126 to 0.705822 (n = 18)
[c] The Voisey's Bay intrusion is hosted by troctolite; genetic association with nearby anorthosite is unclear.
[d] Not true massif-type complexes; both are layered mafic-ultramafic intrusions.

We favour the third hypothesis on the basis of modelling of Re-Os data together with other geologic and geochemical constraints (Hannah & Stein 2002). Nevertheless, the available data sets are limited, and we cannot yet preclude the possibility that real differences exist among the complexes examined to date.

Presently available Re-Os studies of anorthosite-dominated plutonic suites (Table 1) have focused on those with significant sulphide concentrations, including the major ore deposits at Voisey's Bay, Nain plutonic suite, Labrador (Lambert et al. 1999, 2000), and less well-defined deposits in the Suwalki complex, Poland (Stein et al. 1998, Morgan et al. 2000), and the Rogaland complex, Norway (Schiellerup et al. 2000). In each case, initial ^{187}Os-^{188}Os ratios in sulphides are extremely high, indicating a crustal source for the Os. Re-Os data for two additional anorthosite-bearing plutonic complexes also indicate crustal input: the 1845 Ma Sally Malay mafic-ultramafic intrusion in East Kimberly, Western Australia (γ_{Os} of abundant, disseminated sulphides = +950 to +1300; Sproule et al. 1999), and the 1100 Ma Duluth layered igneous complex in Minnesota, USA (γ_{Os} of massive sulphides = +500 to +1200; Ripley et al. 1998).

In two cases, Suwalki and Rogaland, the high initial Os ratios have been attributed to a crustal source for the magmatic complex as a whole (Stein et al. 1998, Morgan et al. 2000, Schiellerup et al. 2000). This hypothesis is supported by a variety of field and mineralogical observations (Taylor et al. 1984, Duchesne 1999), oxygen isotope data (Peck & Valley 2000), and experimental constraints (Longhi et al. 1999). On the other hand, Re-Os data from Suwalki are limited to zones of massive titanomagnetite containing up to a few percent sulphides; modelling demonstrates that observed Re-Os concentrations and isotopic ratios can be explained by selective contamination of a mantle-derived melt by crustal sulphides. In a subsequent and similar study at Rogaland, Schiellerup et al. (2000) call upon a crustal source for the parental magma based on initial $^{187}Os/^{188}Os$ derived from a 10-point Re-Os isochron of 917 ± 22 Ma in agreement with zircon and baddeleyite ages (Schärer et al. 1996). While an isochron initial $^{187}Os/^{188}Os$ of 0.63 ± 0.25 broadly supports this conclusion, the isochron has an extremely high MSWD (517) indicating variation in initial $^{187}Os/^{188}Os$ for the samples analysed. That is, the samples analysed do not form a perfectly coherent genetic suite. Moreover, the data table shows consistent variations in γ_{Os} for individual samples, with sulphides generally higher than oxides. It is conceivable that, as at Suwalki, a larger data set would document selective and variable crustal contamination.

At Voisey's Bay, chondritic γ_{Os} values for silicates lead to a model of selective assimilation of crustal sulphide (and Os) into an otherwise mantle-derived magma

(Lambert et al. 2000). Similarly, both the Duluth and the Sally Malay complexes show contrasts between the isotopic compositions of massive sulphide bodies and minor disseminated sulphide occurrences. The massive sulphides of the Babbitt deposit (Duluth) are strongly crustal, while troctolites with only disseminated sulphides yield epsilon Nd and δ^{34}S of -0.2 to +0.2 and -2.5 to +2.5, respectively (Ripley et al. 1998), typical of chondritic values. Similarly, for Sally Malay, Sproule et al. (1999) report γ_{Os} of +450 to +470 for sulphides from ore-bearing troctolite, and only +60 to +369 for sulphides from weakly mineralized troctolite and peridotite. These data support a model for selective contamination by crustal sulphides, with crustal Os sequestered in an immiscible sulphide melt and not mixing broadly with the complex as a whole.

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Effect of mineralogy and texture of sand and hard-rock ilmenite in TiO$_2$ pigment production by the sulphate process, a case study on Australian ilmenite concentrate and Tellnes ilmenite concentrate, Norway

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The reactivity of the raw material with sulphuric acid is crucial in the production of TiO$_2$ pigments by the sulphate process. Ilmenite concentrates from sand and hard-rock deposits have been studied and the results show that mineralogical characteristics and textural features correlate with the observed recovery rates.

Kemira pigments Oy, Pori, Finland, is treating sand and hard-rock ilmenite concentrates for TiO$_2$ pigment production by the sulphate route. The mineralogy and textural relationships between the various phases determine the reactivity of the concentrate with sulphuric acid, the quality of the pigment and the separability of the phases. The study of commercial ilmenite concentrates provides vital information for the application of sulphate-route pigment manufacturing, by determining the influence on the production processes of the mineralogical characteristics of the raw materials.

Ilmenite concentrate from sand (Australia)

The concentrate is composed of ilmenite and its alteration products, several alteration phases being present in most single grains (Chernet 1999a). Alteration grades from leached ilmenite through pseudorutile to leucoxene. Leached pseudorutile decomposes into leucoxene that consists of aggregates of very fine-grained rutile crystals. The intensity and mode of alteration varies from grain to grain; most commonly, alteration appears along grain boundaries, fractures, crystallographic directions and as leached holes. The concentrate is composed of 53-55% ilmenite plus leached ilmenite, 37% pseudorutile plus leached pseudorutile, and 8-10% leucoxene plus primary rutile. Microprobe analyses of ilmenite and its alteration products show that trace element contents increase with alteration (Chernet 1999a). Ilmenite is soluble in sulphuric acid, whereas rutile is not (Sinha 1979). The degree of solubility of ilmenite, however, depends on the degree of alteration. The concentrate has been separated using a Frantz electromagnetic separator. As the magnetic susceptibility of ilmenite decreases with alteration, the nonmagnetic fractions were recycled in an increasing electromagnetic field. Solubility tests show that solubility decreases with the extent of alteration (Chernet 1999a). The Th content of the concentrate, determined by XRF or ICP, is 0.01%, whereas Ce, Th, Y, La and P$_2$O$_5$ contents in the <90 μm fraction show a 6 to 10-fold increase, due to concentration of monazite into the fine-grained fractions. Uranium contents >3 ppm are only detected in the <90 μm fractions.

Hard-rock ilmenite concentrate (Norway)

The concentrate is composed of 90% ilmenite (hemoilmenite with 8 to 13 mol% hematite), 8% sulphides (pyrite, chalcopyrite, pyrrhotite) and silicates (hypersthene, augite, plagioclase, olivine, biotite), and rare spinel, magnetite, sphene and apatite. Textural features, i.e. exsolution and twinning (Chernet 1999b) are observed. The dissolution of ilmenite and abundance of hematite exsolution were found to correlate. Ilmenite dissolved along grain boundaries with hematite lamellae and as a result hematite lenses are partially leached out and left as oriented pits (Chernet 1999b). Twinning is relatively common in the concentrate but not observed in the corresponding unreacted solid. Cr and V are pigment colorants (Jalava 1992). The concentration of both elements is closely linked to the presence of exsolved hematite (0.25-0.45 wt% Cr_2O_3, 0.4-0.6 wt% V_2O_3). The concentrate contains substantial amounts of ferric iron (32.9 wt% FeO and 12.3 wt% Fe_2O_3), which demands reduction process (Chernet 1999b).

Discussion and conclusions

There is a trend towards an increase in TiO_2 content with decreasing solubility of the concentrate. Advanced alteration products are less soluble than slightly altered ilmenite. Although microfractures and pits favour grinding and percolation of the leachate, the poorly soluble phases are gradually coating these openings, thus progressively preventing the acid to reach the soluble ilmenite. Textural relationships within the different phases and continuous changes of mineral properties and compositions prevent separation of highly altered (poorly soluble) phases from unaltered (soluble) ones. Pigment colouring elements are generally in low contents in the sand concentrates, and the presence of certain amounts of soluble Nb can correct the colour caused by Cr and V. Th and U are concentrated in the fine-grained fraction of the concentrate, mainly as monazite. Discarding the <90 μm fraction considerably reduces their accumulation in the process stream and in the tailings mud.

During the digestion step of the hard-rock ilmenite concentrate, most of the hematite particles will be leached out, and the pits give way to the percolation of the leachate, which resulted in a positive effect on the extraction process. No twinned ilmenite is observed in the corresponding unreacted solid, which indicates that structural discontinuities like twinning and fracturing favour dissolution process. The effect of trace elements on the quality of the pigment is highly dependent on whether it is incorporated in the lattice of the ilmenite or in insoluble phases. Cr and V in this concentrate are closely linked to hematite, which is insoluble.

Knowledge of Ti-bearing phases, including their quantification and alteration products, can throw light on the raw material treatment, enabling the metallurgist and chemical engineers to deal with the problems in TiO_2 recovery. The contents and distribution of trace elements and deleterious minerals provide a useful guide to control the quality of both the final products and the tailings. To conclude, careful mineralogical examination of any ilmenite concentrate is useful in evaluating its suitability for a process.

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Parallel Re-Os isochrons and high ^{187}Os/^{188}Os initial ratios: constraints on the origin of sulphide-bearing anorthosite systems

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In the past decade the Re-Os chronometer and the associated initial ^{187}Os/^{188}Os ratio have provided powerful isotopic tools leading to much improved understanding of the timing and source for ore-related magmas as well as mineralization without a direct magmatic association (Stein et al. 2000, 2001, Arne et al. 2001). Notably, the Re-Os system applied to sulphide- and oxide-bearing anorthosite systems documents unique processes involved in the formation and emplacement of mineralized Proterozoic anorthosite massifs (Hannah & Stein 2002).

Recognition of parallel isochrons and high ^{187}Os/^{188}Os ratios reflecting synchronous emplacement with variable crustal contamination was first intimated at the Proterozoic Suwalki anorthosite complex in Poland (Stein et al. 1998; Morgan et al. 2000). Petrologically, these parallel isochrons reflect complex mingling and addition of crustally-derived Os with anorthosite and related melts. Stoping of crustal rocks by ascending magmas, armoring of crustal conduits, and influx of new batches of melt making their way through new conduits result in highly variable mixtures of crustally-derived and primitive Os (Hannah & Stein 2002). This results in highly variable initial ^{187}Os/^{188}Os ratios in the intruding melts. Because of variable initial ^{187}Os/^{188}Os with different melt batches, it is essential that the resulting Re-Os data be treated on a batch-by-batch basis, and not lumped together on a single isochron. Lumping of data from a large data set may result in an isochron with an approximately correct age, moderate to low uncertainty, but high MSWD. A high MSWD, assuming assignment of appropriate analytical errors, is a clear indication of geological variation in the ^{187}Os/^{188}Os initial ratio and that different magma batches with different Os isotopic histories are represented in the data population. Excess geological scatter resulting

from variable ^{187}Os/^{188}Os initial ratios characterises these data sets (e.g., Schiellerup et al. 2000; Lambert et al. 2000). Correct interpretation of Re-Os data for these mineralized anorthosite systems requires a careful look at the geologic relationships associated with the sample population, and an analysis of the possible balance between crustally-derived Os obtained during the emplacement process and original mantle-derived Os inherent in the anorthosite melt.

Here we show Re-Os isochron plots for two data sets from two different sulphide-bearing anorthosite complexes. In Fig. 1, three different deposits within the Suwalki anorthosite complex in Poland yield Re-Os data that lie on two parallel isochrons with ages of ~1555-1560 Ma (Stein et al. 1998; Morgan et al. 2000). Samples from a single deposit define one isochron, while combined samples from two different deposits define the other. These isochrons are defined by magmatic magnetite (Re = 0.4-1.5 ppb and Os = 0.036-0.144 ppb) and co-existing sulphide (pyrrhotite, chalcopyrite, pyrite, with Re = 30-55 ppb and Os = 1-6 ppb) hosted in anorthosite and related rocks. The ^{187}Os/^{188}Os initial ratios for the two isochrons are 1.16 ± 0.06 and 0.87 ± 0.20, a clear indication of crustal Os and variable initial ratios. In both Figs. 1 and 2, the size of the plotting symbol is more than 100 times the analytical error. The uncertainties reported here are all at the 2-sigma level.

In Fig. 2, the massive sulphides analysed are associated with an anorthosite-bearing, mafic intrusive complex. The locality is not named, as this was a collaborative project with a mining company. The Re-Os data for three samples containing chalcopyrite (cp) and pyrrhotite (po) are shown together with the present-day Os isotopic compositions for mantle and upper continental crust reservoirs. Since
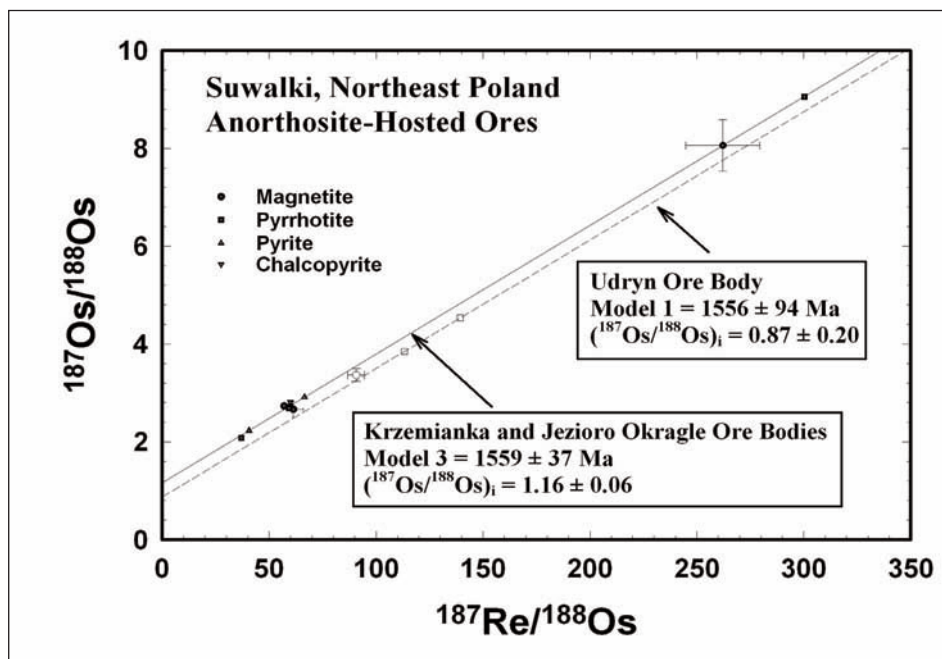
Fig. 1. Re-Os isochrons for magmatic sulphides and oxides from the Suwalki anorthosite massif, northeast Poland (from Morgan et al. 2000).
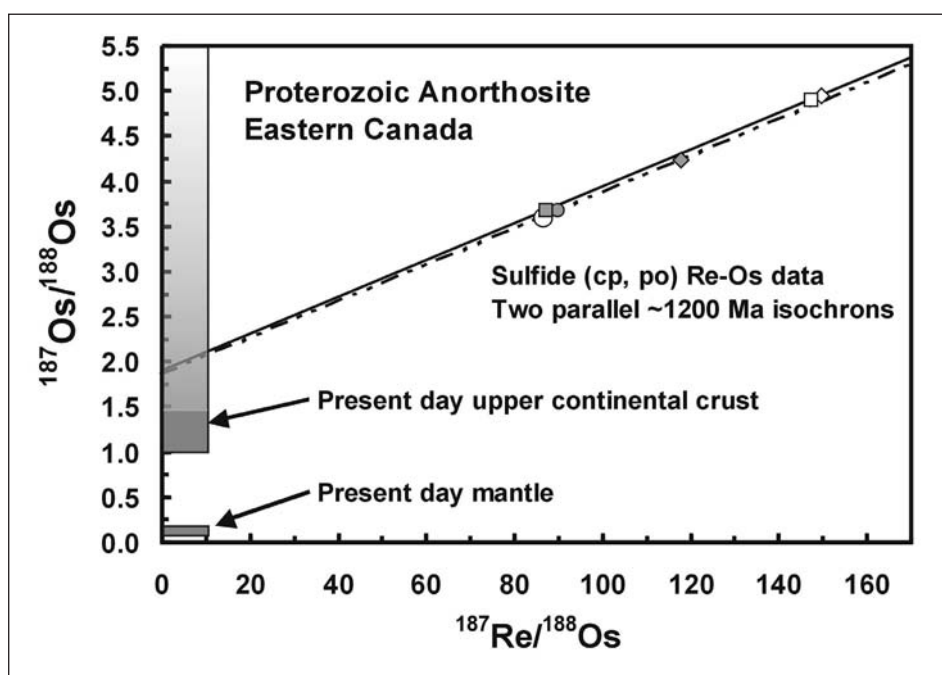
Fig. 2. Re-Os isochrons for sulphides (chalcopyrite and pyrrhotite) related to a Proterozoic anorthosite in eastern Canada. The $^{187}Os/^{188}Os$ isotopic composition for reservoirs representing present-day mantle and average present-day eroding upper continental crust are shown by the dark shaded boxes along the vertical axis. Lighter shading represents $^{187}Os/^{188}Os$ values for other potential crustal contaminants which may extend to greater than 100 (e.g., see Figure 3 in Hannah & Stein 2002).

the $^{187}Os/^{188}Os$ compositions of these two reservoirs (mantle and crust) have evolved (increased) with time, it is important to note that a lower $^{187}Os/^{188}Os$ would have characterised both reservoirs in the geological past. The symbol shapes denote cp and po from the same hand sample; open symbols denote cp and shaded symbols represent po.

For the example in Fig. 2, the Re and Os concentrations are significantly higher than at Suwalki (Re = 29.8-297.0 ppb and Os = 1.51-24.1 ppb). In all three samples, the Os concentrations for pyrrhotite are about twice those of the accompanying chalcopyrite. In two of the samples, pyrrhotite has a higher Re concentration relative to chalcopyrite by about twice, whereas in the third sample, the Re concentrations for both minerals are similar. This suggests that at the hand-specimen scale, pyrrhotite is more effective at taking Re and Os into its structure than chalcopyrite.

If we treat all six analyses together in a regression, we obtain an errorchron of 1239 ± 77 Ma with a very high MSWD = 61 and an initial $^{187}Os/^{188}Os$ of 1.81 ± 0.15. A high MSWD implies excess scatter in the data set, that is, geological scatter in addition to analytical scatter. The most likely source of geological scatter is a variable initial $^{187}Os/^{188}Os$ isotopic composition; that is, the ore 'fluid' composition was variable. There are several ways to create this situation, the most common being mixing from two or more sources (e.g., mixing of magma or magmatic fluid with Os acquired from the country rock, at depth or at the site of emplacement).

A better resolution of the data set in Fig. 2 is obtained using two regressions that yield corresponding ages of 1209 ± 12 Ma and 1201 ± 410 Ma. These regressions are nearly parallel and thus, yield essentially the same age. The very large uncertainty in the 'dashed line' isochron (1201 Ma) is, of course, a function of the fit of points to the line, but also reflects the fact that the spread in these three points is half that for the points on the 'solid line' isochron. What is most striking in these Re-Os results is the indisputably high $^{187}Os/^{188}Os$ initial ratios of 1.908 ± 0.020 ('solid line') and 1.86 ± 0.69 ('dashed line') associated with these sulphides. These very high initial ratios require that the ore 'fluid' had a large component of crustal Os. Also, the crustal component was likely significantly older than 1200 Ma and may have contained phases with high Re/Os ratios; this is necessary to produce highly elevated $^{187}Os/^{188}Os$ ratios by 1200 Ma, the approximate time of contamination of the mafic magma. To add perspective, even the present-day $^{187}Os/^{188}Os$ ratios for 'average crust' cannot explain the data and, as we move back in time to 1200 Ma, the crustal $^{187}Os/^{188}Os$ ratios decrease. Thus, the initial $^{187}Os/^{188}Os$ ratios associated with the samples illustrated in Fig. 2 are extremely high.

We recognize that the construction of multiple parallel isochrons may be necessary to characterise sulphide-bearing anorthosite systems. A single regression for Re-Os data that plots as a band of scattered analyses leads to an approximate or perhaps erroneous age and initial $^{187}Os/^{188}Os$ results. High MSWD values for a regression are a clear indication of geological variation, likely tied to differing initial $^{187}Os/^{188}Os$ ratios in samples. There is much to be learned by coupling the geological relationships with the multiple parallel isochron approach. In particular, precision and accuracy are improved.

It is suggested that the apparent paucity of sulphide in Proterozoic anorthosite systems reflects the required happenstance intersection of ascending anorthosite melt with a crustal, sulphide-bearing rock package. In this way, sulphur with high $^{187}Os/^{188}Os$ may be added to an anorthosite system of primitive derivation through the assimilation or volatilization of older crustal sulphide.

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Fennoscandia

Sveconorwegian rejuvenation of the lower crust in South Norway

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Recent Re-Os isotope data on the c. 0.93 Ga Rogaland Anorthosite Province (RAP) suggest an origin by partial melting of a mafic, lower crustal protolith, without a contribution of mantle-derived Sveconorwegian material (Schiellerup et al. 2000). Without necessarily challenging this conclusion, it should be noted that there is ample evidence of mafic magmatism in South Norway during the Sveconorwegian period (e.g., Munz & Morvik 1991, Haas et al. 2000), which suggests that rejuvenation of the lower continental crust by underplating of mantle-derived magmas could have taken place. The emplacement of the anorthosites and associated rocks of the RAP corresponds in time to a widespread event of post-tectonic, A-type granitic magmatism in SW Scandinavia, which can be used to map the distribution of deep crustal components across this part of the Baltic Shield (Andersen & Knudsen 2000). Sr, Nd and Pb isotope data on late Sveconorwegian granites define mixing trends between crustal components and a component with a juvenile isotopic signature (Andersen et al. 2001a). As pointed out by Andersen (1997) and Schiellerup et al. (2000), the presence of an end member with a positive epsilon Nd does not necessarily indicate that mantle-derived magmas were injected into the crust at the time of deep crustal melting. In fact, such a component may reside in mafic rocks within the deep crust for several hundred million years without totally losing its depleted-mantle Nd isotopic signature.

The hafnium isotope composition of zircon in an igneous rock is a powerful tracer of interaction between the mantle and the lower crust in the past. New laser ablation ICP-MS hafnium isotope data on zircons in Proterozoic granitoids from South Norway (Andersen et al. 2002) allow a better characterization of crustal protoliths in South Norway than has previously been possible, and provide a definitive test for the presence or absence of Sveconorwegian, mantle-derived material in the deep crust of the southwestern part of the Baltic Shield.

In terms of Hf isotope systematics, two distinct pre-Sveconorwegian crustal components can be recognized in South Norway:

i: A Paleoproterozoic component, characterized by depleted-mantle Hf isotope model ages (t_{Hf}^M) > 1.80 Ga. This component resides in the granitoids of the 1.65-1.83 Ga Transscandinavian Igneous Belt (TIB). U-Pb ages and Hf isotope data on

Fig. 1. Cumulative probability plots of time-corrected $^{176}Hf/^{177}Hf$ of zircons from Sveconorwegian granitoids from South Norway, compared to regional hafnium-isotope components. The black parts of the distribution curves represent zircons which cannot have crystallized from a magma formed only from the regional crustal protoliths, and which therefore require the presence of a Sveconorwegian, mantle-derived component in the lower crust.
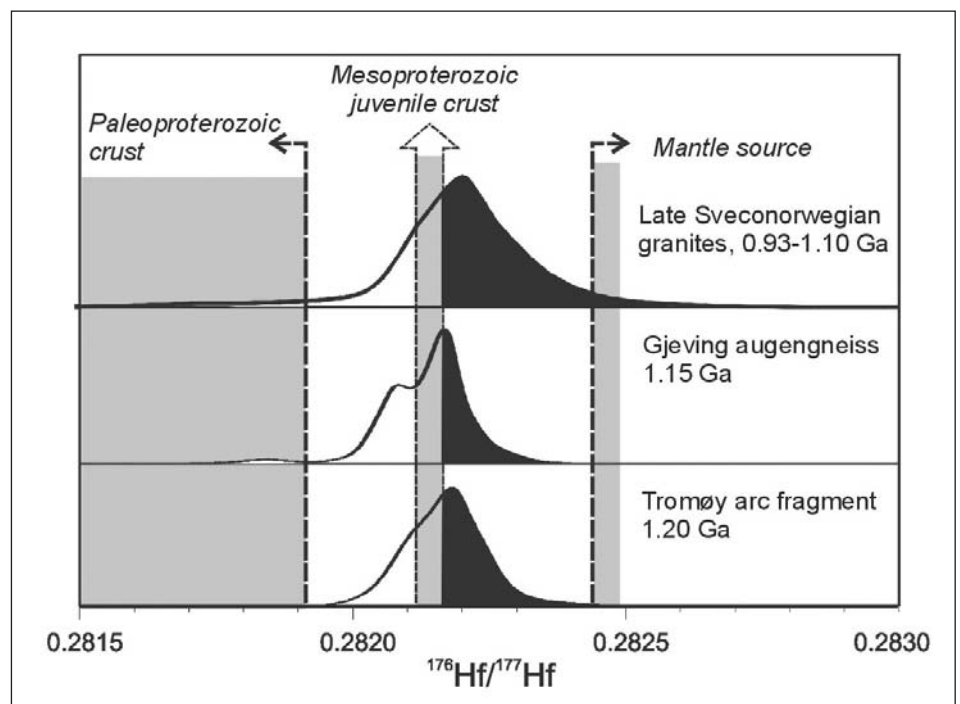
detrital zircons from clastic metasediments indicate that equivalent rocks were present in the source region(s) of Proterozoic sediments in the Telemark and Bamble sectors (Haas et al. 1999). Sr, Nd and Pb data on late Sveconorwegian granites suggest that a similar component can be traced in the deep crust from the area east of the Oslo Rift, across South Norway to the area west of the Mandal-Ustaoset shear zone (Andersen et al. 2001a). In late Sveconorwegian time, the Paleoproterozoic crustal component was characterized by a very distinct hafnium isotope composition, with $^{176}Hf/^{177}Hf \leq 0.2819$ (Fig. 1).

ii: A Mesoproterozoic component with t_{Hf}^M in the range 1.55-1.75 Ga, residing in 1.52-1.60 (-1.66 ?) Ga calc-alkaline, meta-igneous rocks. These rocks are the products of subduction-related magmatism along the SW margin of the Baltic Shield, and can be traced from SW Sweden, across the Oslo Rift and into the southwestern-most part of the Bamble Sector (Andersen et al. 2001b). Both juvenile, mantle-derived material and material with a crustal prehistory were involved in the petrogenesis of these rocks, whose Hf isotope composition in late Sveconorwegian time spans a narrow range at $^{176}Hf/^{177}Hf \approx 0.28215$ (Fig. 1).

A juvenile, depleted-mantle component was involved in the formation of the Mesoproterozoic calca-lkaline rocks, and may later have contributed to the lower crust in one or more underplating event(s). The Hf isotopic composition of a depleted mantle reservoir in Sveconorwegian time can be given by $^{176}Hf/^{177}Hf > 0.28245$ (Fig. 1).

Fig. 1 shows cumulative probability distribution curves of initial $^{176}Hf/^{177}Hf$ ratios of suites of single zircons separated from Sveconorwegian granitoids. All three groups give broad distribution curves, which suggest that the zircons crystallized from isotopically inhomogeneous magmas. Whereas the low-$^{176}Hf/^{177}Hf$ parts of the ranges suggest the involvement of a Paleoproterozoic crustal component, a significant proportion of the zircons from the granitoids show more radiogenic $^{176}Hf/^{177}Hf$ than the Mesoproterozoic crustal component. This indicates the involvement of Sveconorwegian mantle-derived material in the petrogenesis of these rocks.

The Tromøy complex is a fragment of a Sveconorwegian island-arc system, accreted onto the Baltic Shield prior to granulite-facies metamorphism at c. 1.10 Ga (Knudsen & Andersen 1999). The mantle-derived component in these rocks was introduced in the island-arc setting, and does not give evidence of underplating of the continental crust. On the other hand, both the 1.15 Ga augen-gneiss and the 0.93-1.10 Ga granites were emplaced within the continent, and the presence of a significant component of mantle-derived hafnium in their zircons is a clear indication that the granitic magmas did not form by simple remelting of a Mesoproterozoic deep crustal protolith. Input of mantle-derived material to the lower crust during the Sveconorwegian period is thus necessary to account for the initial Hf isotope characteristics of these rocks.

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Thermal modeling of the metamorphism of the Rogaland granulites

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The Proterozoic Egersund Anorthosite Complex is composed of anorthosite and related rock types, which intrude intercalated charnockitic and garnetiferous migmatites. The highest metamorphic temperatures are spatially associated with the intrusion, and previous workers recognized high-temperature mineral isograds for pigeonite-in, osumilite-in and orthopyroxene-in. The P-T conditions at a distance of 16 km from the contact are about 700°C and 5 kbar and increase to more than 1000°C and 5 kbar at 2.5 km.

The high-temperature metamorphism in the country rocks found near the contacts of the northeastern part of the Rogaland intrusive complex (surface area of about 1000 square km) cannot be explained by assuming the heat source was a simple, single phase of intrusion. In order to obtain the observed metamorphic temperatures and isograd distribution, thermal modeling indicates that the heat source must have had at least two main phases that were separated by a hiatus of about 3 - 4.5 m.y. In this model, emplacement and crystallization of the anorthosite (areal extent: 30 x 40 km) produces a thermal gradient (750–600°C) in the country rocks. While this thermal gradient is developing, a second, smaller intrusion (Bjerkreim-Sokndal lopolith, areal extent: 9 x 12 km)

is emplaced at the anorthosite-country rock contact. Because the country rocks nearest the anorthosite have undergone appreciable heating, the second intrusion can provide enough thermal input to

Fig. 1. The diagram depicts the thermal model of the northeastern part of the Rogaland intrusive complex, and the distribution of maximum temperatures and duration isochrons, which indicate the total amount of time that the rock spent above 750°C. See text for discussion.
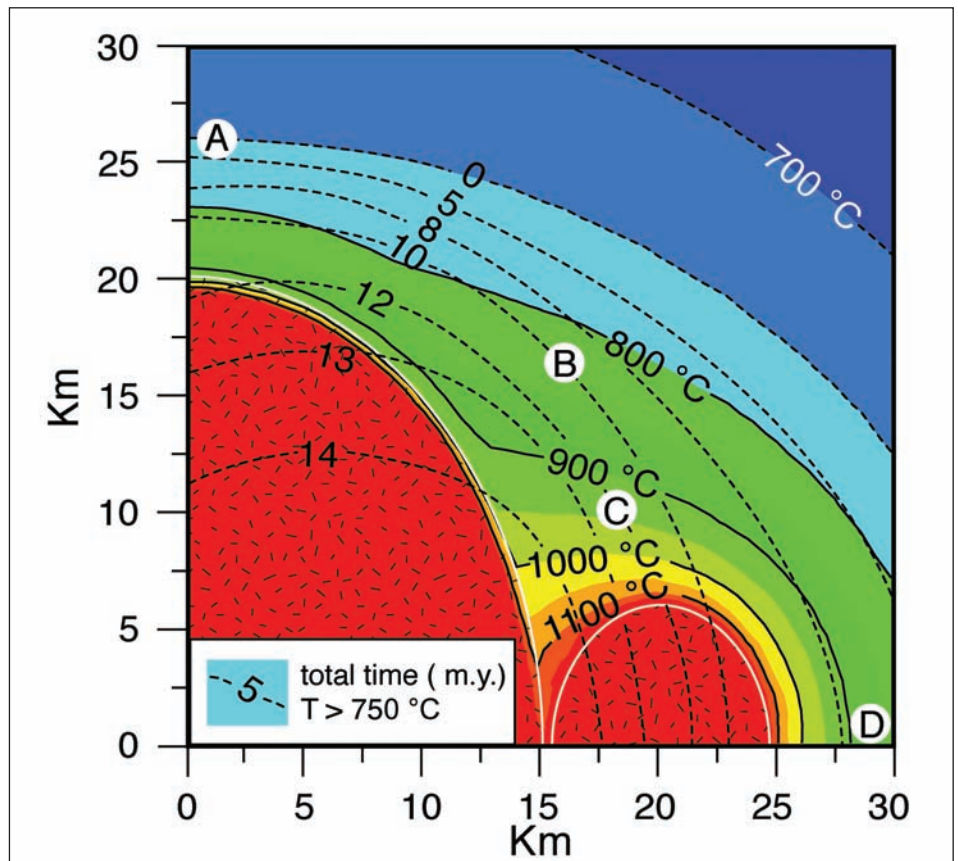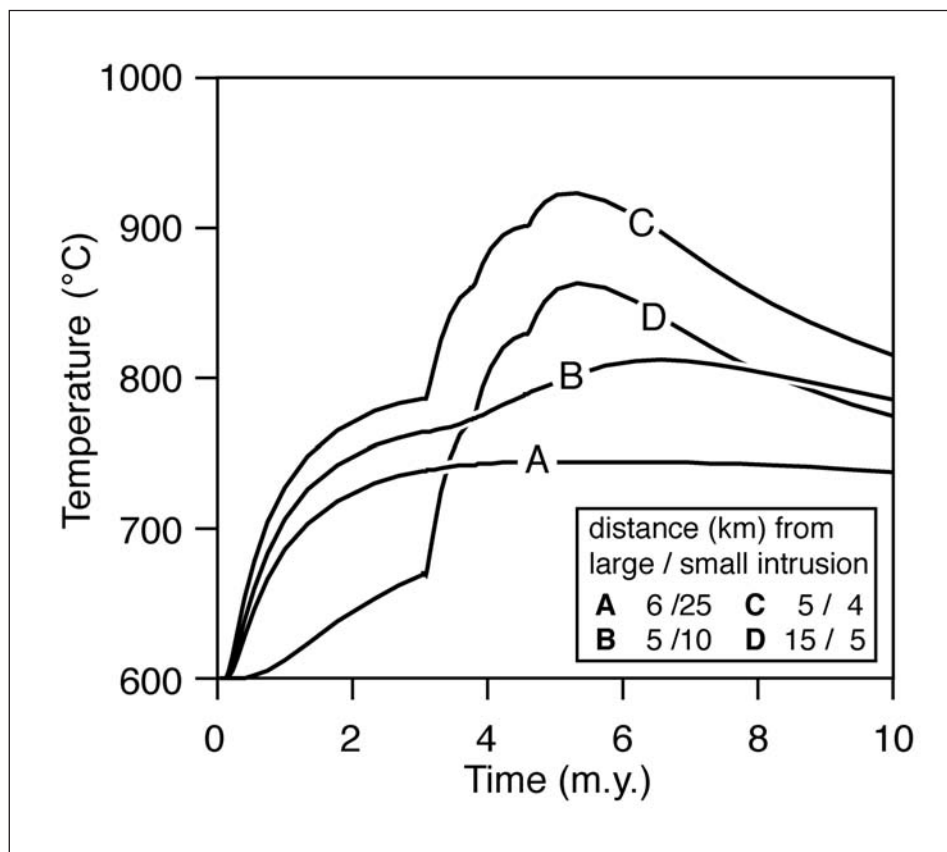
Fig. 2. The diagram shows the variation in temperature versus time based on the thermal modeling experiment. The four curves are for the points (A-D) on Figure 1.

obtain the observed high temperatures. Recent work (Schärer et al. 1996) indicates that the entire magmatic emplacement occurred over a time interval of about 10 m.y. (930-920 Ma), which is consistent with the thermal model of the metamorphism.

From the thermal modeling experiment, we have estimated the distribution of maximum temperatures and duration isochrons, which represent the length of time the rocks spent above 750°C during heating and cooling (Fig. 1). The duration isochrons roughly parallel the contact of the initial and larger intrusion, while the maximum temperature isotherms are strongly deflected by the smaller and later intrusion. This indicates that, while the second intrusion has a major impact on the local temperature distribution, it does not locally prolong the heating and cooling cycle. The 750°C isotherm and the duration isochon for 0-5 m.y are roughly parallel over the entire area. The duration isochrons of 10 to 14 m.y. crosscut the isotherms at steep angles near the smaller and later intrusion (near point C in Fig. 1), while near point A, at greater distances from the smaller intrusion, the isotherms and duration isochrons are nearly parallel.

These relative orientations of the isotherms and the duration isochrons show the interaction of the two intrusions. The effects are strongest in the area between the intrusions (near point C in Fig. 1), and the effect of the smaller intrusion dissipates noticeably at distances greater than 10 km (near point B in Fig. 1). The area around point C (Fig. 1), where the isotherms and duration isochrons are most oblique, gives an impression of the extensive variation in heating and cooling rates that can occur over relatively small areas.

Figure 2 shows the variation of temperature with time for the four lettered points in Figure 1. Point A is located 5 km from the intrusive contact along the extension of the long axis of the large, elliptical intrusion. Point B is located at 5 km distance to the large and 10 km to the small intrusion and lies on 10 m.y. duration isochrons. The initial rapid heating and long period of cooling for these localities (A and B) are controlled by the anorthosite (Fig. 2). The difference in the temperatures at points A and B after 3 m.y. reflects the geometrical effect (elliptical shape) of the modeled intrusion (see also point C).

For point C the heating and cooling history is radically different from points A and B (Fig. 2). The variations in heating and cooling rates and the higher maximum temperature are due to the major influence of the second intrusion (modeled BSL).

At point D the thermal effects of the second intrusion on the country rocks (heating and cooling rates) are more drastic than at point C, but the maximum temperature is lower because of the lower initial temperatures at the time of the second intrusion. The major differences between the thermal development at points C and D (Fig. 2) are due to differences in country-rock temperature at the time of the second intrusion (i.e., the magnitude of the initial thermal gradient caused by the anorthosite).

The P-T estimates from 5 and 10 km from the intrusive contact and data from the thermal model were also used to estimate cooling rates and time intervals. Maximum temperatures are 860°C and 800°C and temperatures greater than 750°C (apparent end of new garnet growth) were maintained for 4.5 to 8.5 m.y. About 5 km

from the intrusion, calculated post-peak cooling rates indicate a decrease from 30° to 10°/m.y. over about 7 m.y., while at about 10 km calculated post-peak cooling rates decrease from 12° to 7°/m.y. over about 7 m.y. During this time interval, retrograde and fine-grained Gar-Qz rims formed around orthopyroxene and primary garnet (outermost 150-200 µm) present in assemblages of garnet-quartz-plagioclase-orthopyroxene ± spinel. Based on the widths of preserved zoning profiles in garnet and orthopyroxene and diffusion data, estimates of cooling rates can be made, and these agree well with the calculated temperatures and cooling rates from the model. Where garnet and orthopyroxene are in direct contact, the retrograde exchange may have continued as long as 15-25 m.y., which resulted in more highly developed Fe/Mg-zoning at the rims (outermost 40-240 µm) of both orthopyroxene and garnet.

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Re-Os dating of the Ørsdalen W–Mo district in Rogaland, S Norway, and its relationship to Sveconorwegian high-grade metamorphic events

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In the Rogaland–Vest Agder sector of the Sveconorwegian orogen (S Norway), a granulite-facies metamorphic domain is exposed around the 931 ± 3 Ma Rogaland Anorthosite Province (Duchesne et al. 1985, Schärer et al. 1996). This domain is characterized by a succession of three main isograds with increasing grade towards the west: the orthopyroxene isograd in leucocratic rocks, the osumilite isograd in paragneiss, and the pigeonite isograd in leucocratic rocks (Tobi et al. 1985). The osumilite and pigeonite isograds are indicative of very high-temperature and dry conditions. Osumilite-bearing assemblages provide P–T estimates of 5.5 kbar–800-850°C (Holland et al. 1996). There is little doubt that osumilite-bearing assemblages developed as a result of granulite-facies contact metamorphism related to the intrusion of the anorthosite complex (M2 metamorphism), as the osumilite isograd is parallel to the external contact of the plutonic complex. Available petrological and geochronological data on monazite in the region nevertheless point to high-grade metamorphism, possibly in granulite-facies conditions, before the intrusion of the anorthosite complex (M1 metamorphism; Tobi et al. 1985, Bingen & van Breemen 1998). In this work, we use the Re-Os chronometer applied to molybdenite hosted in small metamorphic deposits, to further explore the geochronology of polyphase metamorphism in Rogaland.

Reliable and precise ages for molybdenite crystallization can be obtained by the Re–Os method using improved analytical methods (Markey et al. 1998, Stein et al. 2001). The Re–Os chronometer applied to molybdenite has been shown to maintain its systematics through high-grade metamorphism (Stein et al. 1998, Raith & Stein 2000). The method is thus suitable for obtaining primary ages of molybdenite deposition, regardless of metamorphic overprinting. Among the small molybdenite-bearing deposits in Rogaland, the Ørsdalen district was mined for wolframite, scheelite, and molybdenite between 1904 and 1956. It is located between the orthopyroxene and osumilite isograds. It occurs in a granitic to granodioritic gneiss unit that commonly contains metamor-

phic garnet and/or orthopyroxene. The mineralized zone is restricted to two ca. 50 m thick layers. These layers are composed of biotite gneiss with amphibolite lenses and contain glassy quartz + feldspar leucocratic veins up to 1 m wide, both elongate parallel to the gneiss foliation (Heier 1955, 1956, Urban 1974, Olerud 1980). The W-Mo mineralization is closely associated with the quartz-rich veins.

Samples were collected at the Mjåvassknuten (MJ1 and MJ2) and Stopulen (ST1) prospects, situated within the two mineralized layers. Sample MJ1 is taken from irregular, coarse-grained, glassy quartz + plagioclase + orthoclase leucocratic masses that are associated with a fine-grained and dark-coloured granulite, rich in brown-red biotite. Both the leucocratic masses and the host contain disseminated coarse-grained dark red garnet and orthopyroxene. Molybdenite occurs as clots and patches larger than 1 cm that are made of fine- to coarse-grained crystals. It is dominantly hosted in the leucocratic masses and associated with scheelite. Molybdenite, biotite and Fe-Ti oxide minerals are commonly surrounded by a fine-grained symplectite of garnet + quartz. Samples MJ2 and ST1 are from concordant, clear, glassy quartz-rich veins in biotite gneiss. The molybdenite is concentrated along vein margins as mm- to cm-sized blebs to stringers.

The Re-Os data for the three molybdenite samples are presented in Fig. 1 with one replicate analysis for MJ2. All model ages include the uncertainty in the ^{187}Re decay constant and overlap at the 2-sigma level of uncertainty. Their weighted mean age is 971.4 ± 3.6 Ma. The four data points provide a Model 3 isochron age (Ludwig 1999) of 973.3 ± 8.2 Ma with the expected zero initial ^{187}Os intercept of -0.10 ± 0.88 (2-sigma). Inclusion of the zero point, implicit in Re–Os model ages for molybdenite and as demonstrated in the four-point regression, provides essentially the same Model 3 age of 972.9 ± 3.8 Ma. The five-point regression contributes to the reduced uncertainty.

The agreement of four Re-Os ages using three different samples is a clear verification of a robust Re-Os chronometer in molybdenite, and points to an event of deposition of molybdenite at about 973

Ma. This age is younger than any possible deposition age or magmatic crystallization age for the amphibolite lenses associated with the deposit. A presumably Mesoproterozoic stratiform volcanogenic origin for the deposit, as proposed by Urban (1974) is thus not supported by this new data. The field and petrographic relations in the Mjåvassknuten locality indicate that the leucocratic molybdenite, garnet- and orthopyroxene-bearing quartz + feldspar masses represent veins and pods of partial melt formed during an event of granulite-facies metamorphism. By direct inference, we propose that the molybdenite age of 973 Ma is directly related to the formation of the leucocratic veins and pods. This event is related to the granulite-facies metamorphic and deformation event that produced coarse-grained metamorphic assemblages, produced a strong foliation in the country rocks, produced leucocratic veins and lenses, and aligned these veins within this foliation. The occurrence of garnet + quartz symplectite around molybdenite shows that the deposit was affected by an event of static high-grade metamorphism after deposition (M2 contact metamorphism at about 931 Ma). The Re–Os systematics in molybdenite were not affected by this subsequent event.

The age of 973 Ma for molybdenite deposition lies within the range of proposed timings for Sveconorwegian metamorphism in the orogen. In the easternmost sector of the orogen in SW Sweden, a high-pressure granulite-facies domain with local occurrence of eclogite-facies rocks is well constrained between 980 and 960 Ma (Johansson et al. 2001). This metamorphism probably resulted from thrusting of the allochthonous terranes onto the foreland of the orogen. In the westernmost sector of the orogen, in Rogaland, single-grain monazite U–Pb ages define a range between 1024 and 970 Ma for M1 regional metamorphism and between 930 and 925 Ma for M2 contact metamorphism (Bingen & van Breemen 1998). The occurrence of negatively discordant monazite grains at 970 Ma in some biotite-bearing gneiss samples suggests that the age of 970 Ma is geologically significant. The new data on the Ørsdalen district suggest formation of mineralized quartz + feldspar granulite-facies veins at 973 Ma, and thus give independent evidence that M1 regio-

nal metamorphism reached granulite-facies conditions at 973 Ma. Available data suggest a protracted nature for M1 metamorphism between ca. 1025 and 970 Ma. They also point to the superposition of two granulite-facies metamorphic events in Rogaland (M1 and M2 events), and that osumilite-bearing mineral associations formed during the second of the two events. Superposition of two granulite-facies events may be required to sufficiently dehydrate a gneissic basement so that osumilite-bearing assemblages are developed and preserved. The necessity for two superimposed stages of granulite-facies metamorphism may explain the rarity of osumilite-bearing assemblages in granulite-facies domains.

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Tectonic setting of the Egersund-Ogna anorthosite, Rogaland, Norway, and the position of Fennoscandia in the Late Proterozoic

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A detailed magnetic study has been undertaken on the Egersund-Ogna anorthosite body of the Rogaland Anorthosite Province, southeastern Norway. The anorthosite, with published U-Pb ages of 930 Ma, is part of a larger complex of massif-type anorthosites and the Bjerkreim-Sokndal layered intrusion. Thirteen paleomagnetic sites were collected in the body, representing various parts of the anorthosite. Average susceptibilities range from 0.03 to 2.24 x 10-3 SI and NRM intensities range from 0.004 to 1.54 A/m. Corresponding Q values range from 3 to 148 with a mean value of 36, indicating remanent-controlled magnetic anomalies. Remanent directions from all samples are characterized by steep negative inclinations with southwest to northeast variable declinations. Thermal demagnetization reveals square shouldered demagnetization curves, with little or no loss of intensity until 550 or 575° C. Alternating field demagnetization produces a wide range of demagnetization behaviors with mean destructive fields varying from less than 5 mT to greater than 80 mT. There is little evidence of overprinting or secondary components, and all information points to a remanence gained during initial cooling of the anorthosite. Mean directional data for the 13 sites are I = -81.7° and D = 326.8°, a95 = 6.0. Assuming this mean direction represents normal polarity, paleolatitude for southern Fennoscandia at this time is 70°S. The magnetic pole calculated for Egersund-Ogna is, at -44° latitude and 198° longitude, in good agreement with earlier poles determined from other Rogaland Anorthosite Province rocks. This work supports reconstructions that place Baltica in high (southern) latitude at approximately 900 Ma. Apparent polar wander paths for Baltica at this time are ambiguous with both clockwise or counter-clockwise loops proposed. These data confirm the presence of mid-latitude poles in the Late Proterozoic, but at this point are unable to discern between suggested scenarios.

Sulphides in the Rogaland Anorthosite Province

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The Rogaland Anorthosite Province in Southwest Norway is a composite massif-type anorthosite province consisting of a number of anorthositic plutons, mafic, dominantly jotunitic or noritic intrusions, as well as a collection of granitoid bodies. Many mafic intrusions form significant deposits of Fe-Ti oxides.

Sulphide occurrences in Rogaland may be divided into two types. 1: minor disseminated, lumpy or semi-massive occurrences within anorthosite bodies, and 2: disseminated sulphides hosted by noritic or orthopyroxenitic cumulates believed to derive from jotunitic magmas. Type 2 occurrences include a number of noritic intrusions that contain dispersed sulphide droplets, but only in one instance did a concentration process accompany the sul-
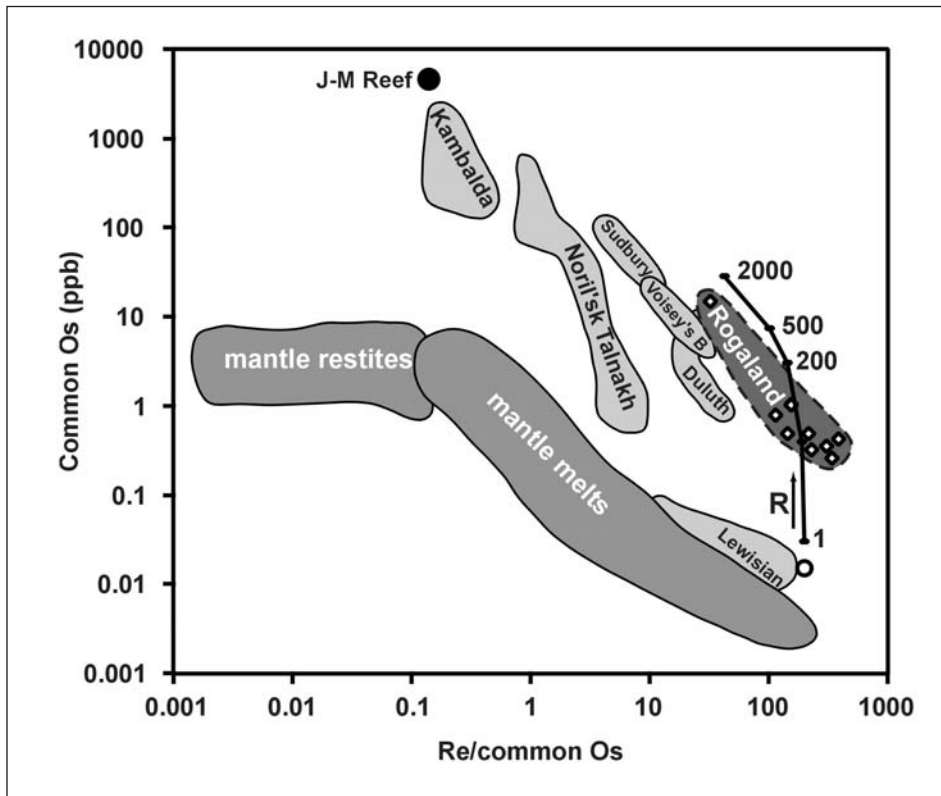
Fig. 1. Common Os (Os concentration at the time of formation) vs. Re/Os ratio in Rogaland sulphides compared with major magmatic sulphide deposits and mantle melt compositions. Diagram and global data fields adopted from Lambert et al. (1999). All sulphide data have been normalized to 100 % sulphide. The inferred parental magma composition is shown with an open circle.

phide saturation, to form a significantly sulphide-enriched cumulate.

Type 1 deposits are hosted by anorthosite plutons, but usually occur in spatial association with slightly more evolved noritic or even pyroxenitic cumulate pockets, pegmatite dykes or intrusions. The sulphide occurrences are not confined to marginal sites, but may be found scattered throughout the anorthosite bodies. Re-Os and other chalcophile element modelling (Fig. 1) suggest that all anorthosite-hosted deposits, except one, are characterised by very low R-factors (ratio between amounts of silicate melt in effective equilibrium with the sulphide melt). The low R-factors (~50) imply that little interaction between silicate and exsolved sulphide melt took place. The scattered occurrence, low R-factors and association with lower-temperature cumulates, imply that the sulphide melt exsolved from the interstitial silicate melt at a late stage in the anorthosite formation. Elevated R-factors (~500) in anorthosite-hosted occurrences are only observed in the Fossfjellet deposit located in the deformed margin of the Egersund-Ogna anorthosite.

Sulphide droplets are ubiquitous in all major mafic intrusions in Rogaland, which must have been near-saturated in sulphur prior to emplacement. Fractional segregation generally dispersed sulphide melt droplets evenly in the mafic cumulates, and only in the Bjerkreim-Sokndal Layered Intrusion did sulfides segregate to form a significant stratiform type 2 deposit in a noritic intrusion. An elevated R-factor (3-500) is required to explain the composition of this occurrence.

Isochronous relationships among Fe-Ti rich norites, anorthosites and contained sulphides have been recorded by Re-Os and Sm-Nd isotopes, that yield isochron ages of 917±22 and 914±35 Ma, respectively (Fig. 2). These ages strongly corroborate existing U-Pb dating in the province (Schärer et al., 1996), and the existence of isochrons is good evidence that the source characteristics for the mafic intrusions, anorthosites and sulphides are similar. Rare marginal chills, correlated mineral chemistry, restricted initial Sr isotope ratios and phase equilibrium considerations imply that all mafic rocks are cumulates formed at various stages during the differentiation of jotunitic magmas. In addition, experimental studies have recently shown that the Rogaland anorthosite bodies may also be derived from jotunitic magmas (Longhi et al., 1999). R-factor processes and modal variations are indeed able to account for all compositional vari-

Fig. 2. Regional Re-Os and Sm-Nd isochron diagrams. 10 Re-Os samples form a model 3 isochron with an elevated MSWD of 519. Data from the Storgangen and Tellnes intrusions have been reset and form a poorly defined trend close to the Caledonian 425 Ma reference line. 30 Sm-Nd samples yield a model 3 isochron with an MSWD of 27. Initial ratios correspond to γ_{Os} = +419 and ε_{Nd} = +1.47.

that sulphide saturation in the anorthosites and mafic intrusions resulted from the differentiation of silicate magmas of comparable compositions. No external source for sulphur seems to be required.

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ability among the sulphides, whether hosted by anorthosite or mafic intrusions.

The late sulphide saturation in the anorthosites, the appearance of sulphides in lower-temperature mafic intrusions, and the compositional similarities, suggest

Fe-Ti-V-P deposits in anorthosite complexes: the bearing of parental magma composition and crystallization conditions on the economic value

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Recent experimental data (Fram & Longhi 1992, Longhi et al. 1993, Longhi et al. 1999, Vander Auwera et al. 1998) indicate that parental magmas of the AMC suite probably encompass a large continuum of compositions ranging from high-Al basalts to more ferron and potassic compositions, represented by jotunites (hypersthene monzodiorites). Experimental phase equilibria show that both end member magmas can account for the norite series which fractionates at 3-5 kb to silica-enriched liquids (Longhi et al. 1999).

In Rogaland, comparison between phases experimentally obtained on a primitive jotunite (Vander Auwera & Longhi 1994)

and natural phases from the anorthosite massifs and from the Bjerkreim-Sokndal layered intrusion suggests that a liquid generally similar to this jotunite was parental to both types of intrusions. Interestingly, in these two types of intrusions, Fe-Ti-V-P deposits have been recognized (Duchesne 1999). In massive anorthosites, the ore-bodies occur as (deformed) dykes or pods ranging in composition from pure ilmenite (Jerneld) to ilmenite norite (Tellnes, Storgangen). Polybaric fractional crystallisation and syn-emplacement deformation in rising anorthosite diapirs lead to relatively (poisonous) Mg- and Cr-rich ilmenite (± V-magnetite) deposits. Fractional crystallization in laye-

red magma chambers of jotunite magmas gives rise to voluminous 'disseminated' mineralization, containing low Mg and Cr ilmenite + Ti-magnetite ± REE-rich apatite, still of sub-economic value. Immiscibility is not the controlling mechanism, except perhaps in some rare nelsonites (Hesnes).

Subsolidus re-equilibration leads to a thorough change in the oxide mineral composition towards an enrichment in end-member compositions (Duchesne 1970). Reactions between the oxide minerals leave conspicuous microscopical evidence and regularly lower the hematite content and possibly the Cr content of ilmenite, and the ilmenite and Al-spinel contents of the magnetite. Mg contents of ilmenite also seem to decrease, possibly through reactions with pyroxenes, but no reaction rims can be observed at the contact with pyroxene.

Experimental data suggest that there is little pressure dependence on the oxide mineral compositions (Vander Auwera et al. 2001). Therefore, irrespective of possible subsolidus re-adjustment, the main controlling factor on the primary ilmenite composition (and value) is the composition of the melt in which it crystallized. The latter, in turn depends on the parental magma composition and on the degree of fractional crystallization that the melt attained when the deposit formed. Further complexity can be introduced by the fact that the parent magmas of jotunitic composition display a wide range of trace element and isotope contents (Duchesne et al. 1989).

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Whole-rock geochemistry of the Bjerkreim-Sokndal layered series: bearing on crystallization processes of cumulus rocks

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Crystallization processes of cumulate rocks from the Bjerkreim-Sokndal layered intrusion (Rogaland, southwest Norway) (Wilson et al. 1996) have been investigated through a whole-rock major and trace elements geochemical approach. More than 100 whole rocks from the layered series of the Bjerkreim lobe (MCU II to IV) have been analysed for major elements. Two types of cumulates are particularly well represented in the collection: (1) phi-C or leuconorite cumulate: plag (An_{50} to An_{45}) + hemo-ilmenite + opx (± olivine ± magnetite); (2) phimac-C or gabbronorite cumulate: plag (An_{45}-An_{36}) + opx + cpx + ilmenite + magnetite + apatite (Duchesne 1978).

In Harker variation diagrams (Fig.1), the two types of cumulates show well-defined linear trends for each element vs. SiO_2. The regression coefficients are generally excellent: from 0.92 to 0.96 in the leuconorites, from 0.89 to 0.97 (except for Ca, with r = 0.53) in the gabbronorites. A simple linear 2-component mixing model can thus account for the major element compositions. The first component is plagioclase alone (An47 in leuconorite cumulate, and An40 in gabbronorite cumulate), and the second one is made up of the sum of the other minerals, i.e. the mafics (hemo-ilmenite and opx for leuconorite cumulate; ilmenite, magnetite, opx, cpx and apatite for gabbronorite). In other words, as a first approximation, any cumulate rock composition oscillates between an anorthosite pole and a mafic pole. Small deviations from the linear area can be attributed to analytical errors, to the presence of minor amounts of trapped liquid in the cumulate, or to the effect of cryptic layering. Large deviations occur when the simple model is not respected, but such cases are very rare (5 cases out of 105).

This particular property has three important implications: (1) the relative proportion of the different mafic minerals (pyroxenes, oxide and apatite) remains constant during each specific stage of evolution; (2) though cryptic layering is conspicuous in the layered series (Duchesne 1978, Wilson et al. 1996) and particularly in each type of cumulates compositional variations in individual minerals tend to perfectly balance each other, so that the plagioclase and mafic poles remain globally constant; (3) the variation of the whole-rock compositions along the linear trend only results from variation in the relative abundance of plagioclase in the cumulate.

A purely gravity-controlled mechanism of accumulation of minerals at the bottom of the magma chamber, controlled by the Stokes' law, is likely to modify the relative proportions of the mafic minerals because of the large spread in mineral densities (from 5.2 for magnetite to 3.2 for apatite). Furthermore, the density difference between plagioclase and apatite is lower than the density contrast between apatite and ilmenite or magnetite, though the proportion of the latter minerals remains constant. A process of in-situ crystallization of the plagioclase has already been suggested because of the small density difference between the plagioclase and the melt (Duchesne & Hertogen 1988, Vander Auwera et al. 1994). The present data suggest that all minerals have formed in situ, and that oscillatory nucleation of plagioclase alone is responsible for the variation of the plagioclase content.

Fig. 1. Harker diagrams showing the linear trend in the two types of cumulates: black open circle: leuconorite cumulates (phi-C); red cross: gabbronorite cumulates (phimac-C); blue filled square: mafic pole for leuconorite; red filled diamond: mafic pole for gabbronorites; the dashed line connect the mafic poles to the average plagioclases: green triangle: An_{40}; inverted blue triangle: An_{47}.
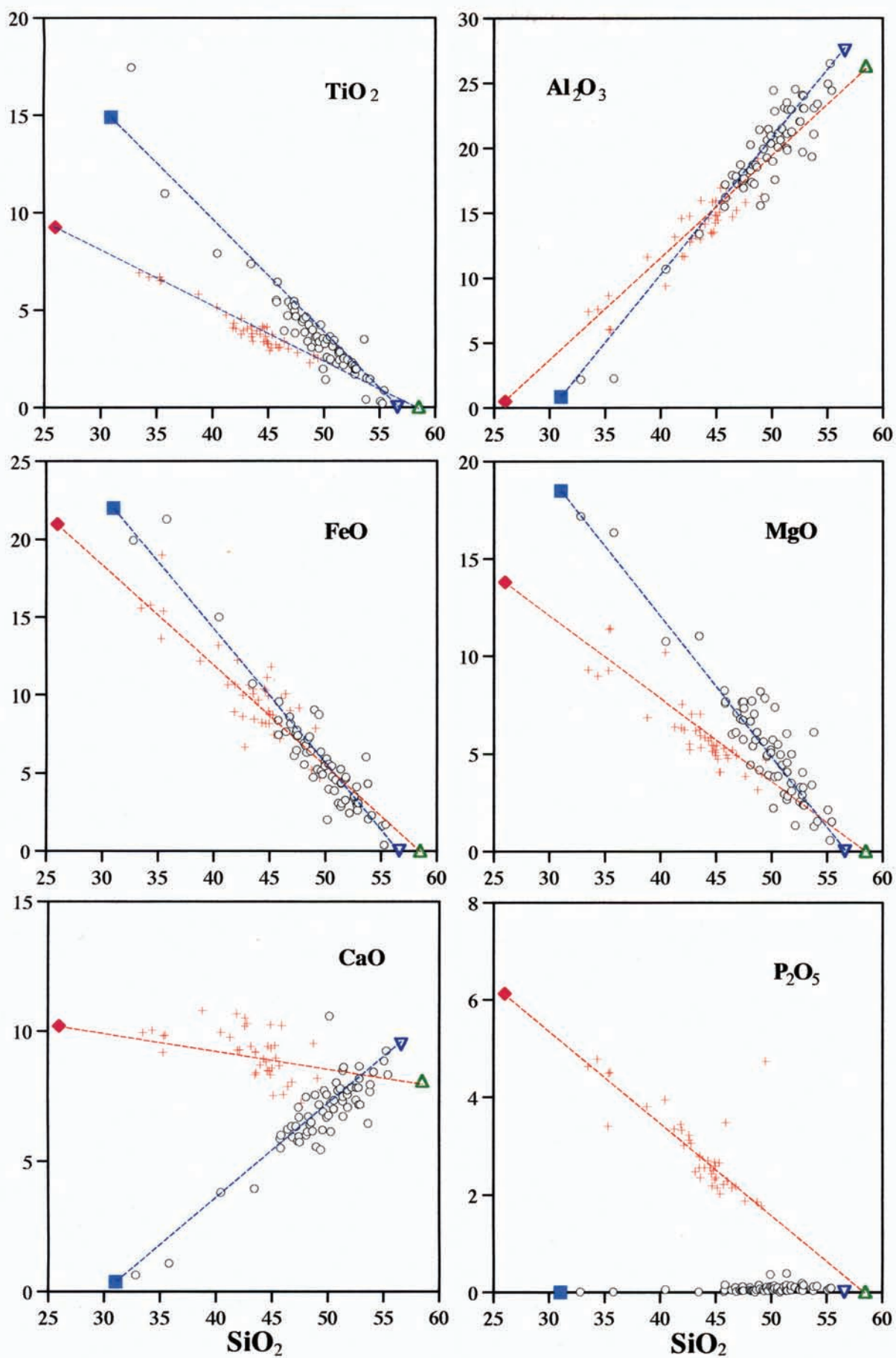
Forty selected rocks were also analysed for trace elements. REE distributions clearly discriminate the two types of cumulates. Leuconorite cumulates are poor in REE and show a large positive Eu anomaly (Eu-Eu*= 3 to 10), whereas gabbronorites contain 8 to 10 times more REE and present flat normalized patterns without significant Eu anomalies. The REE (except Eu) content of leuconorite is proportional to the amount of P_2O_5, which itself reflects the abundance of trapped liquid. The flat normalized patterns of gabbronorites implies that the positive Eu anomaly of the plagioclase is swamped by the negative anomalies of the mafics, and particularly by that of apatite. REE analyses of minerals show that cumulus apatite has indeed a small negative Eu anomaly, and REE contents are 100 to 300 times more abundant than plagioclase. Cumulus apatite thus buffers the REE content and the Eu/Eu* ratio of the cumulate, irrespective of the other mineral contents and, particularly, of the plagioclase content.

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Relationship between the layered series and the overlying evolved rocks in the Bjerkreim-Sokndal intrusion, southern Norway

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The Bjerkreim-Sokndal layered intrusion consists of a >7000 m-thick Layered Series comprising anorthosites, leuconorites, troctolites, norites, gabbronorites and jotunites (hypersthene monzodiorites), overlain by an unknown thickness of massive, evolved rocks (mangerites (MG), quartz mangerites (QMG) and charnockites (CH)). The Layered Series is subdivided into 6 megacycic units (MCUs) that represent the crystallisation products of successive major influxes of magma. The relationship between the Layered Series and the overlying, evolved rocks has long been debated. Are the two parts of the intrusion comagmatic or do the evolved rocks represent a separate magma influx?

We have studied a c. 1200 m-thick profile that straddles the sequence from gabbro-norites belonging to the top of the uppermost MCU to the quartz mangerites in the northern part of the intrusion (the Bjerkreim lobe). Mineral compositions in 37 samples become continuously more evolved in the lower part of the sequence up to the middle of the MG unit (plagioclase An_{36-18}; olivine Fo_{39-7}; Ca-poor pyroxene mg# 57-16; Ca-rich pyroxene mg# 65-22). Above this, compositions are essentially constant in the upper part of the MG unit and in the QMG (An_{20-13}; Fo_{6-4}; Mg#opx17-13; Mg#cpx25-18). There is no evidence of a compositional break that could reflect a magma influx event in this profile.

Whole-rock chemical compositions also show a continuous evolutionary trend. The MG unit defines a cumulate trend

essentially controlled by the amount of alkali-feldspar (which is mostly mesoperthite). All of the QMG samples lie in a small field with e.g. 0.8-0.3% MgO and 59-65%SiO_2. Duchesne & Wilmart (1997) identified two compositionally distinct types of QMG-CH in the upper part of the intrusion. One type (olivine-bearing QMG and CH) defines an olivine trend (OT) which they interpret essentially as reflecting fractional crystallisation of the mangerite liquid resident in the chamber. The other (two-pyroxene QMG and amphibole CH) defines the 'main liquid line of descent' derived from a jotunitic liquid.

The entry of quartz as a cumulus phase, a major concentration of country rock inclusions (the 'septum xenolithique') and

two-pyroxene QMG all occur at the same stratigraphic level. This could reflect disruption of the roof related to an influx of jotunite-derived magma from an external source, as suggested by Duchesne & Wilmart (1997). An alternative is that the two-pyroxene QMG and amphibole CH represent a roof melt and that convective overturn with the underlying, evolved, resident magma resulted in collapse of the roof.

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SEM elemental mapping and characterisation of ilmenite, magnetite and apatite from the Bjerkreim-Sokndal layered intrusion

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The grade and grain-size distribution of ore minerals are key factors in evaluating economic mineral deposits. This abstract reports the results of a mineral characterisation based on SEM elemental mapping and image processing. The work is part of the Geological Survey of Norway's ongoing exploration for apatite, ilmenite and vanadium-bearing magnetite in the Bjerkreim-Sokndal layered intrusion.

The samples for this study were collected by several co-workers (Korneliussen et al. 2001, Meyer & Mansfeld, this volume). The stratigraphical and petrogenetic interpretations of some of the samples are by Meyer & Mansfeld (this volume). Selected thin-sections were measured by SEM using EDS XRF-RAY elemental mapping. With our SEM instrumentation, we obtain picture files for up to 16 different elements (one picture for each element). The images are recorded with a magnification of 16 times and the mapped area covers about 1/3 of a standard thin-secti-
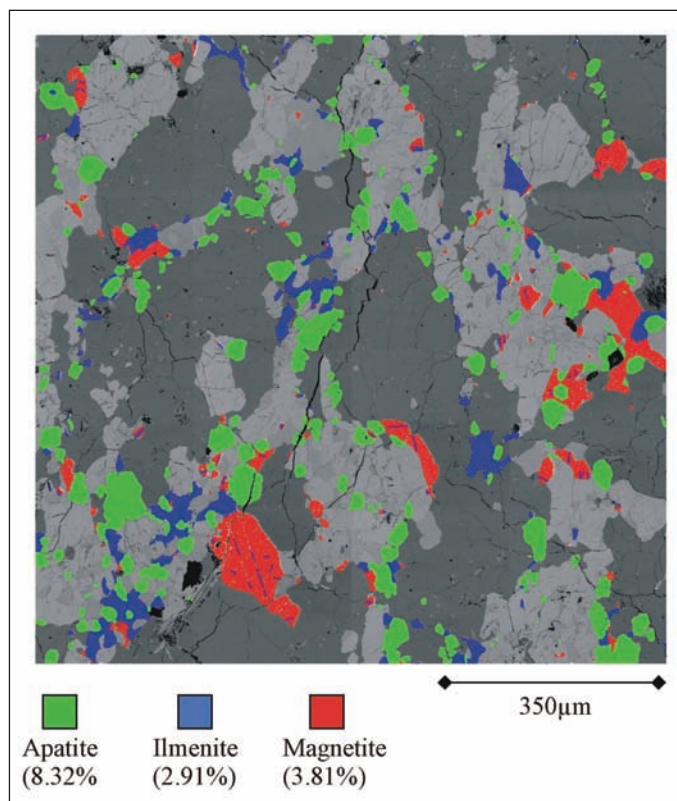
Fig. 1. Example of coloured backscatter image, with modal content of apatite, ilmenite and magnetite.

350μm

| Apatite (8.32%) | Ilmenite (2.91%) | Magnetite (3.81%) |
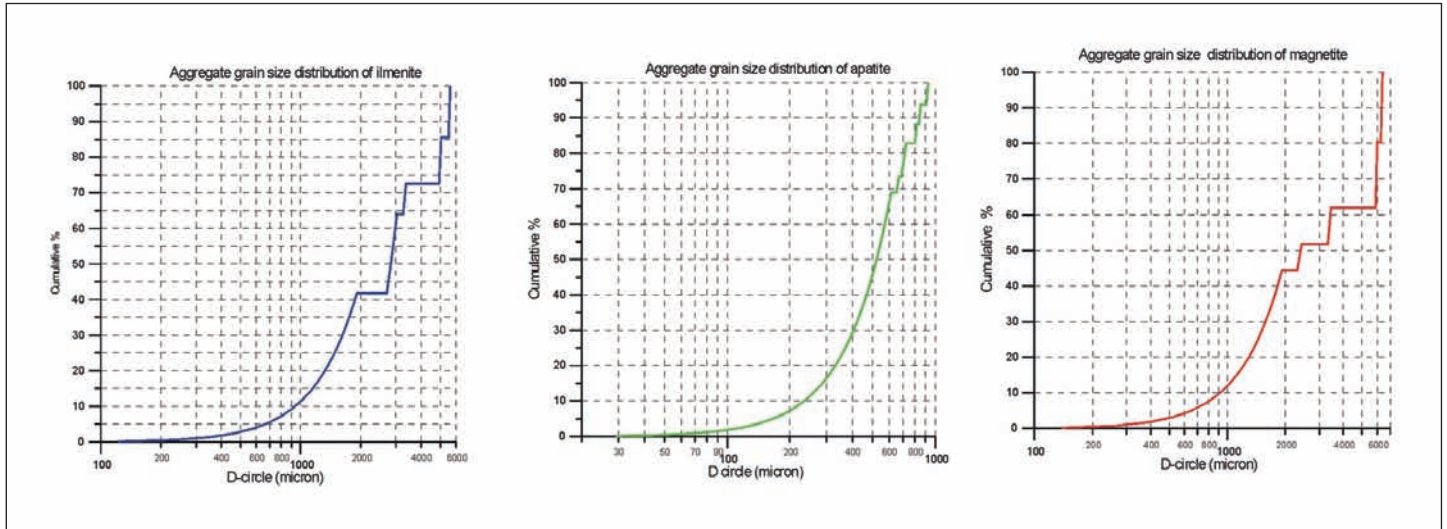
Fig. 2. Grain-size distribution (aggregate data from all samples in Table 1) for apatite, ilmenite and magnetite measured by SEM imaging and processing. (d-circle is the diameter of a circle with the same area as the measured particle).

on. The x-ray element maps were processed using the KS300 image processing software (a commercial program from Carl Zeiss vision). The X-ray images were converted to binary images, which again were improved by morphological operations (several steps combining erosion and dilation) to remove noise, artificial grain contacts, etc. Finally, by some steps of *boolean* image combination, new images were generated which represent the area distribution of selected minerals.

The image processing software KS300 automatically measures the morphological features (such as area, longest and shortest axis, axis of enclosing ellipse of each mineral grain, etc.). In addition, the total modal % of all minerals in the rock and the modal % of ilmenite, magnetite and apatite within selected size fractions are calculated (Fig. 1). The grain-size distribution is shown in Fig. 2. Table 1 shows the total content of ilmenite, magnetite and apatite and the content within selected grain-size classes in the 35 studied samples.

Our analysed samples have the following modal and statistical data:

Ilmenite: The average mode is 14.17% and the maximum value 54.71%. 2.7% of the total ilmenite in a sample is on average less than 50 μm . The D_{50} grain size (d-circle) is 2850 μm in diameter.

Magnetite: The average mode is 11.44% and the maximum content 51.95%. 4.50% of the total magnetite is on average less than 50 μm. The D_{50} d-circle length is 2400 μm.

Apatite: The average mode is 5.74% and the maximum content 12.78%. 16% of the samples total apatite is in grains on average less than 50 μm. The D_{50} d-circle length is 523 μm.

The sample (K323a) that contains the highest content of apatite (12.78%) also contains 11.46 % of ilmenite and 6.56% of magnetite. This is a sample from the mega-cyclic unit III, north of Vasshus.

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Table 1. Content of apatite, ilmenite and magnetite in Bjerkeim-Sokndal rocks (unit refers to the megacyclic units of the intrusion)

Sample.	Locality	Coordinates (utm east-north)	Unit	% apatite in size class (microns)							% Ilmenite in size classes (microns)							% magnetite in size classes (microns)						
				% total ap.	%ap <50	%ap 50-100	%ap 100-150	%ap 150-200	%ap 200-250	%ap >250	% total ilm.	% ilm <50	% ilm 50-100	% ilm 100-150	% ilm 150-200	% ilm 200-250	% ilm >250	% total mt	% mt <50	% mt 50-100	% mt 100-150	% mt 150-200	% mt 200-250	% mt >250
BJ9901	Orrestad	337150	6493625	MCUIV	7.22	0.04	0.19	0.4	0.82	0.79	4.97	5.12	0.1	0.23	0.24	0.4	0.74	3.43	1.77	0.16	0.21	0.17	0.1	0	1.12
BJ9902	Orrestad	337200	6493750	MCUIV	8.15	0.05	0.24	0.71	1.11	0.72	5.32	4.65	0.1	0.1	0.29	0.2	0.24	3.69	4.64	0.04	0.11	0.16	0.31	0.19	3.82
BJ9903	Orrestad	337100	6493850	MCUIV	10.72	0.06	0.16	0.39	0.69	1.13	8.29	4.07	0.14	0.15	0.23	0.2	0.56	2.85	3.01	0.1	0.28	0.17	0.38	0.09	2
BJ9905	Orrestad	337200	6494000	MCUIV	4.52	0.05	0.14	0.3	0.43	0.39	3.2	1.96	0.16	0.1	0.14	0.2	0.15	1.26	1.96	0.16	0.1	0.14	0.16	0.15	1.26
BJ9907	Orrestad	337050	6493425	MCUIV	5.28	0.02	0.1	0.28	0.6	0.46	3.82	4.44	0.13	0.15	0.33	0.3	0.14	3.33	4.72	0.12	0.29	0.23	0.18	0.31	3.58
BJ9909	Orrestad	337200	6493250	MCUIV	8.32	0.11	0.65	1.07	1.35	0.93	4.22	2.91	0.34	0.29	0.36	0.2	0.64	1.12	2.91	0.34	0.29	0.36	0.17	0.64	1.12
BJ9910	Orrestad	337100	6493150	MCUIV	6.39	0.04	0.17	0.58	0.81	0.66	4.14	4.71	0.33	0.17	0.1	0.2	0.19	3.75	6.86	0.22	0.32	0.36	0.23	0.16	5.56
BJ9912	Orrestad	338100	6492950	MCUIV	8.41	0.06	0.24	0.58	0.68	0.99	5.86	13.4	0.22	0.21	0.15	0.2	0	12.64	6.21	0.34	0.32	0.16	0.08	0.44	4.87
K320a	Svalestad	333560	6491370	MCUIII	10.62	0.42	1.05	1.56	1.99	1.85	3.75	9.05	0.30	0.35	0.76	0.78	1.22	5.64	9.74	0.14	0.50	0.53	0.45	0.71	7.41
K320b	Svalestad	333560	6491370	MCUIII	11.28	0.41	0.91	1.28	1.33	0.89	6.46	6.11	0.15	0.26	0.19	0.3	0.56	4.7	13.7	0.2	0.37	0.31	0.33	0.32	12.2
K322a	Vasshus	346100	6491000	MCUIII	11.35	0.61	1.99	2.37	2.58	2.47	1.33	8.28	0.48	0.78	0.99	1.1	1.2	3.77	10.61	0.2	0.65	0.66	0.74	0.88	7.47
K323a	Vasshus	346110	6491050	MCUIII	12.78	0.22	0.91	1.4	1.8	1.77	6.7	11.5	0.26	0.32	0.46	0.5	0.39	9.52	6.56	0.37	0.36	0.44	0.35	0.47	4.56
K324a	Orrestad	336490	6492810	MCUIV	11.97	0.15	0.81	1.16	1.39	1.22	7.24	9.2	0.07	0.14	0.38	0.5	0.32	7.79	9.8	0.18	0.5	0.4	0.55	0.4	7.77
LPN088	Bakka	343350	6473890	MCUIV	10.15	0.32	1.24	1.32	1.24	1.47	4.57	16.5	0.07	0.09	0.2	0.5	0.49	15.18	14.9	0.08	0.37	0.55	0.6	0.67	12.6
LPN085	Bakka	343390	6473790	MCUIV	11.24	0.38	1.49	2.77	1.67	1.93	2.99	18.9	0.03	0.16	0.27	0.2	0.51	17.75	23.58	0.13	0.56	0.56	0.41	0.56	21.4
LPN090	Mydland	349610	6475800	MCUIV	7.36	0.21	0.59	0.8	0.85	1.08	3.82	7.32	0.02	0.07	0.15	0.2	0.14	6.76	9.69	0.11	0.26	0.34	0.29	0.32	8.36
LPN144	Hauge	3410909	6470010	MCUIV	0.27	0.09	0.17	0.02	0	0	0	54.7	0.31	0.16	0.07	0	0	54.17	31.39	0.05	0.13	0.04	0.04	0.34	30.8
LPN148	Årstad 01	342710	6469020	MCUIV	0	0	0	0	0	0	0	27.6	0.46	0.22	0.16	0.2	0.26	26.32	64.3	0.06	0.06	0.02	0.04	0.1	64
LPN166	Sokndal	340480	6469630	MCUIV	0.06	0	0.03	0	0	0	0	51.3	0.01	0.08	0.09	0.2	0.25	50.62	0	0	0	0	0	0	0
LPN167	Sokndal	341270	6469030	MCUIV	0.04	0.04	0	0	0	0	0	18	0.08	0.2	0.31	0.4	0.78	16.25	8.39	0.07	0.07	0.22	0.19	0.43	7.42
LPN184	Sokndal	341050	6475540	MCUIV	0.04	0.04	0	0	0	0	0	24.3	0.13	0.37	0.63	0.7	1.01	21.55	6.05	0.14	0.19	0.4	0.45	1.2	3.66
LPN187	Sokndal	342390	6472010	MCUIV	0.07	0.03	0.01	0	0.04	0	0	20.6	0.18	0.12	0.15	0.2	0.15	19.85	51.95	0.09	0.09	0.08	0.06	0.1	51.5
LPN19219	Sokndal	343140	6472760	MCUIV	1.67	0.2	0.46	0.59	0.24	0.07	0.12	31.8	0.07	0.07	0.12	0	0	31.53	24.9	0.06	0.17	0.22	0.34	0.14	24
LPN19223	Sokndal	343140	6472960	MCUIV	2.42	0.25	0.73	0.6	0.46	0.24	0.14	25	0.09	0.2	0.13	0.1	0.34	24.07	29.92	0.39	0.33	0.33	0.41	0.67	27.8
LPN223	Mydland	343180	6473960	MCUIV	2.14	0.17	0.27	0.38	0.49	0.24	0.59	16.1	0.1	0.29	0.43	0.9	0.99	13.47	1.81	0.19	0.29	0.24	0.25	0.49	0.35
LPN227	Åmdal	346200	6477050	MCUIV	9.22	0.16	0.97	1.57	1.93	1.55	3.03	11.7	0.13	0.25	0.08	0.2	0.25	10.79	4.6	0.24	0.22	0.27	0.55	0.33	3
LPN231	Rekefjord	340000	6468950	Rekfj.	5.3	0.13	0.75	1.1	0.98	0.4	1.94	3.98	0.07	0.19	0.56	0.3	0.44	2.47	2.27	0.04	0.12	0.13	0.04	0.37	1.57
LPN236	Eia	341500	6481750	Rekfj.	2.48	0.33	0.92	0.67	0.32	0.23	0	2.04	0.12	0.3	0.31	0.5	0.07	0.78	1.94	0.24	0.25	0.34	0.4	0.29	0.42
LPN238	Teksevatn	339830	6493640	MCIII	8.25	0.32	1.16	1.82	2.04	1.4	1.51	8.66	0.21	0.65	0.83	1	0.74	5.28	2.52	0.23	0.45	0.42	0.13	0.39	0.91
LPN248	Eigeland	350700	6466700	Anorth.	2.11	0.81	0.91	0.33	0	0.07	0	2.6	0.39	0.54	0.36	0.3	0.08	0.93	8.57	0.29	1.18	1.51	1.2	0.88	3.52
LPN252	Dragland	340200	6471900	MCUIV	0.37	0.08	0.08	0.06	0.06	0.08	0	24.4	0.08	0.24	0.53	0.6	1.22	21.8	3.61	0.05	0.08	0.19	0.31	0.32	2.66
LPN255	Bilstad	344250	6491200	MCUIV	0.65	0.14	0.17	0.09	0.14	0	0.1	3.47	0.18	0.45	0.68	0.7	0.38	1.11	5.66	0.5	0.71	0.65	0.37	0.06	3.37
LPN256	Bilstad	345050	649100	MCUIV	10.91	0.31	0.95	1.42	1.96	1.42	4.85	9.02	0.21	0.44	0.61	0.6	0.54	6.66	0	0	0	0	0	0	0
LPN260	Bakka	342450	6472150	MCUIV	9.19	0.28	1.13	1.45	1.67	1.64	3.02	14.4	0.08	0.33	0.46	0.2	0.38	13.02	13.61	0.16	0.23	0.22	0.33	0.37	12.3
Averages				**5.74**	**0.19**	**0.56**	**0.77**	**0.85**	**0.75**	**2.63**	**14.14**	**0.17**	**0.25**	**0.34**	**0.37**	**0.46**	**12.57**	**11.44**	**0.16**	**0.29**	**0.32**	**0.30**	**0.38**	**9.99**

LA-HR-ICP-MS studies of ilmenite in MCU IV of the Bjerkreim-Sokndal intrusion, SW Norway

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The layered Bjerkreim-Sokndal Intrusion forms part of the Rogaland Anorthosite Province of South Norway (Wilson et al. 1996). The province is known for its voluminous anorthosite bodies and ilmenite-rich noritic intrusions of which the Tellnes Dyke contributes to c. 6% of the present world Ti-production. Earlier studies of ilmenite compositions in the noritic bodies were primarily based on analyses by electron microprobe or by XRF-analyses on ilmenite concentrates (e.g. Duchesne 1972). These studies have given significant knowledge of the ilmenite compositions in many of the noritic intrusions in the province. However, the need for high-precision in situ analyses of ilmenite has grown during the last few years. Based on this need we have developed a method for analysing ilmenite by Laser Ablation High Resolution Inductive Coupled Plasma-Mass Spectrometer (LA-HR-ICP-MS). Here we present the method based on 24 samples from a profile covering a 2400 m-thick cumulate sequence. The profile is situated in the northern part of the Bjerkreim-Sokndal Intrusion.

The northern part of the Bjerkreim-Sokndal Intrusion consists of a c. 7 km-thick Layered Series, which is divided into megacyclic units (MCU) arranged in a syncline. Each MCU formed in response to magma replenishment in a bowl-shaped
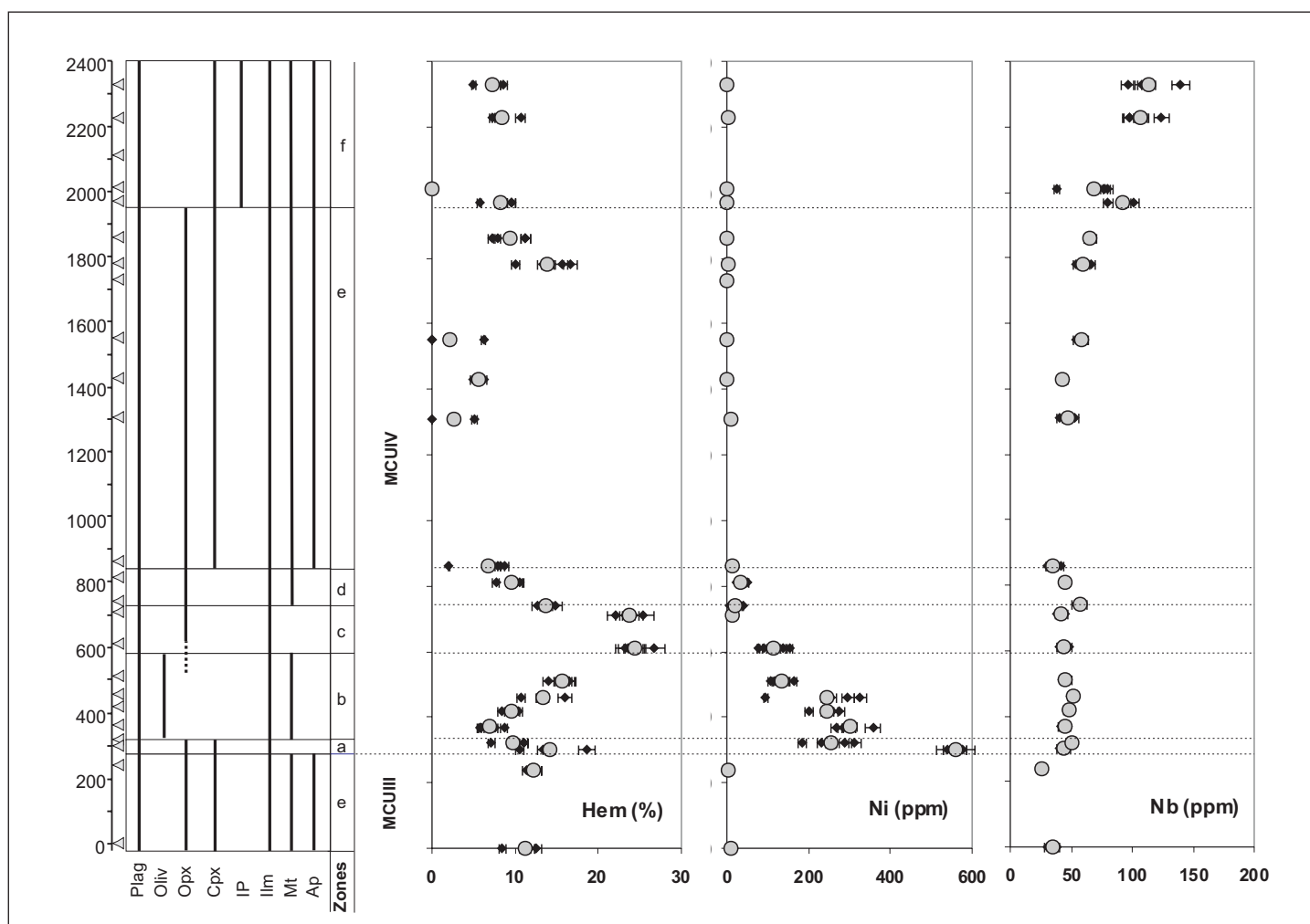
magma chamber. The megacyclic units are characterised by common crystallisation sequences of which the fourth and last (MCU IV) is the most complete. The profile we present here covers the uppermost part of MCU III and a major portion of MCU IV (Fig. 1).

The samples were analysed by UV-laser ablation in medium resolution (M/ΔM-=3000) on a Finnigan MAT ELEMENT® double focusing sector ICP-MS. Previous to the ablation, the ilmenite was analysed by electron microprobe. Continuous laser ablation of the samples was performed as a 100 by 100 µm raster, giving the bulk composition of ilmenite plus hematite exsolution lamella in each grain. The elements Ti, Fe, Mg, Mn, Si, Ca, Al, V, Cr, Ni, Zr and Nb were analysed. The concentration of the major elements (Ti, Fe, Mg and Mn) was determined by stoichiometric calculations in combination with element calibration. Trace elements were calibrated against international and synthetic ilmenite and rutile standards that were manufactured from $TiFeO_3$ powders doped with trace elements in varying concentrations. The standard powders were melted by direct fusion in graphite electrodes resulting in fast melting and quenching. Two or more analyses of 10 different standards were used to establish the calibration curves, which allowed outliers caused by inhomogeneous standards or analytical noise to be excluded. The analytical and calibration uncertainty is between 2-5% in total and the detection limits are 0.1-10 ppm for trace metals and less than 0.1 % for major oxides. Parts of the results are presented in Fig.1 together with

the cumulate stratigraphy for the Terland profile.

In the upper part of MCU III, ilmenite contains about 10% hematite, increasing slightly across the MCU III/ IV boundary to 13%. In zone a of MCU IV, the hematite content gradually falls to 7 %, and then in zone b and c increases gradually to 25%. Above this level the hematite content gradually decreases to 5-10% in the central and uppermost part of the profile. Ni shows an increase in concentration at the lowermost level of MCU IV (zone a), and then a gradual decrease to zone d, where Ni reaches a minimum of a few ppm, corresponding to an increasing degree of fractional crystallisation. Nb shows classic incompatible element behaviour in terms of increasing concentration in response to increasing degree of fractional crystallisation. The trends described are in good agreement with the trends described in earlier works for the silicate minerals (Jensen et al. 1993), which indicate a major replenishment event of magma reflected by the reversal in composition evolution in the lower part of MCU IV, followed by fractional crystallisation of magma throughout MCU IV. However, the hematite component seems to be strongly controlled by the onset and offset of other crystallising mineral phases such as magnetite, olivine and Ca-poor pyroxene (opx).

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Fig. 1. Stratigraphic profile through MCU IV and the uppermost part of MCU III from the Terland area. The cumulate stratigraphy is shown on the left, where the profile is divided into zones based on terminology by Wilson et al. (1996). Triangles along the y-axis (stratigraphic height in metres) indicate each sample locations. On the right, the results from the Laser-ICP-MS analyses of ilmenite are shown for hematite (Hem (%)), for Ni (ppm) and for Nb (ppm). Circles represent average compositions for each sample, while filled diamonds represent individual analyses within each of four grains for each sample.

Geochemistry and mineralogy of the Tellnes ilmenite ore body

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The Tellnes ilmenite ore body, one of the largest ilmenite deposits of the world, has been mined by Titania A/S since 1960. The ore body, which is of magmatic origin, appears as a sickle-shaped lens about 2700 m long and 400 m wide in the Åna-Sira anorthosite massif, belonging to the Rogaland Anorthosite Province.

Whole-rock geochemical analyses have been carried out on drill cores from the ore body since the initial prospecting of the mine in 1955. At present, the geochemical database consists of about 5000 analyses including TiO_2, P_2O_5, S, Cr and modal magnetite. Furthermore, more than 2000 samples have been analysed for the remaining major elements and a selection of trace elements (Fig. 1). In addition, mineral compositional data have been analysed from a selection of drill cores.

The ore is characterised by the lack of modal layering. However, the ore body shows a strong compositional zoning. The TiO_2 content is generally high (>18 wt %) within the central parts of the ore body, and there is a gradual decrease in the TiO_2 content towards the contact to the host-rock. The variations in the ore chemistry reflect the variations in the modal contents of the ore-forming minerals. Based on these variations, the ore body has been divided in four zones (Fig. 1): the upper marginal zone (UMZ), the upper central zone (UCZ), the lower central zone (LCZ), and the lower marginal zone (LMZ). It should be noted that this division is approximate, because the transitions between the various zones are smooth. The UMZ is characterised by high modal contents of plagioclase and Fe-Mg silicates (clinopyroxene, orthopyroxene and olivine), but low contents of ilmenite. The UCZ shows high modal contents of ilmenite and Fe-Mg silicates, and low contents of plagioclase. The LCZ, which is the most ilmenite-enriched zone, is characterised by low contents of Fe-Mg silicates (olivine is absent), and intermediate contents of plagioclase. Through the LMZ, the ilmenite content decreases, while the contents of plagioclase, clinopyroxene and orthopyroxene increase.

The ore-forming minerals show compositional variations that are highly correlated to the whole-rock chemistry (Fig. 2). In general, the Mg/(Mg+Fe)-ratios of ilmenite and the Fe-Mg silicates are high in the central TiO_2-rich parts of the ore body (UCZ and LCZ). Towards the margins, however, Mg/(Mg+Fe)-ratios of the ore-forming minerals decrease gradually.
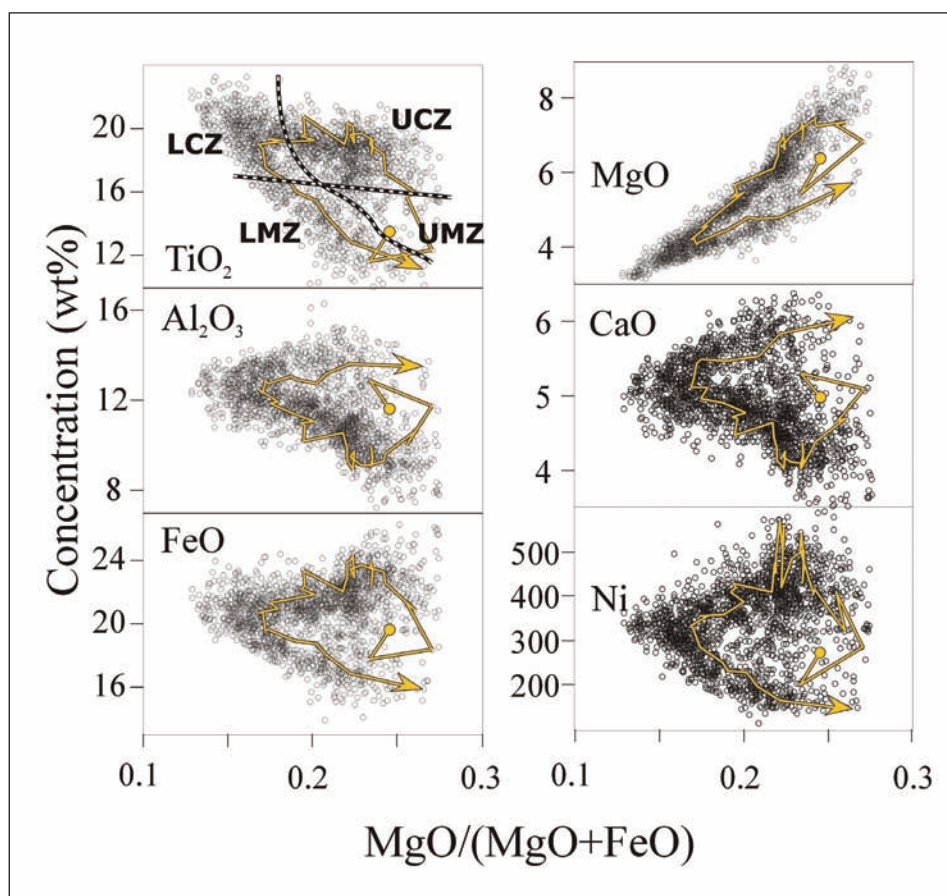
Fig. 1. Selected components of the ore vs. MgO/(MgO+FeO). The compositional variations of drill core 9.5V-2N (grey shaded line) from top (circle) to bottom (arrow) have been indicated. The white dashed lines in the upper left diagram separate the upper marginal zone (UMZ), the upper central zone (UCZ), the lower central zone (LCZ), and the lower marginal zone (LMZ). Note the gradual changes in composition along drill core 9.5V-2N from the upper marginal zone of the ore body, through the central part, and finally, through the lower marginal zone.
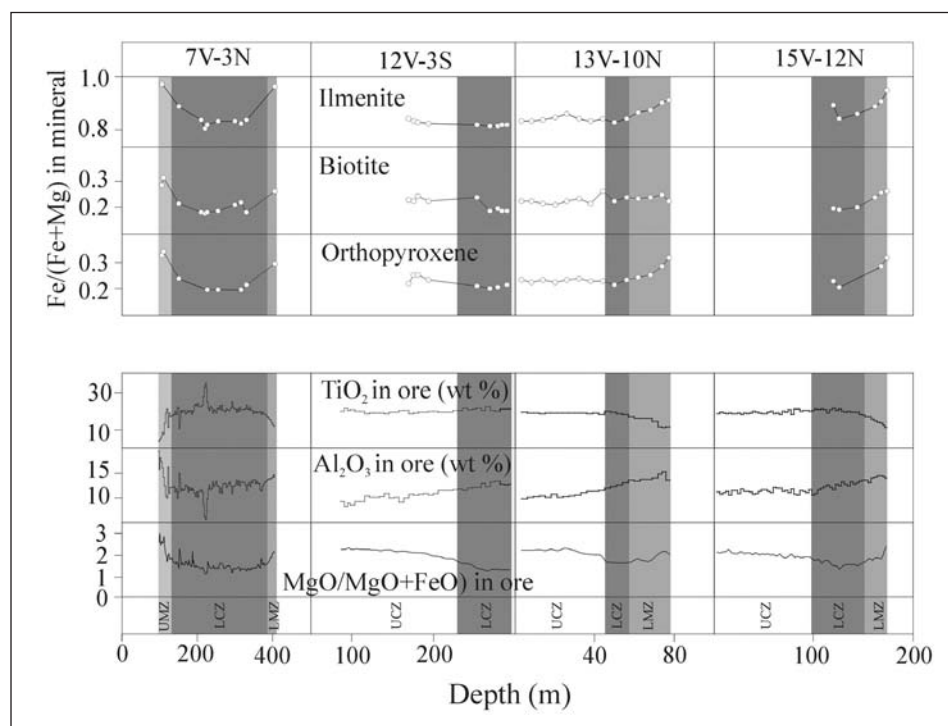
Fig. 2. Variations of selected chemical components of minerals and whole rocks along drill cores 7V-3N, 12V-3S, 13V-10N and 15V-12N. Note the gradual compositional changes through the ore body, with respect both to mineral chemistry and to whole-rock chemistry. The Fe/(Fe+Mg) ratios of the minerals are generally high along the upper and lower margins of the ore body, but low in the central parts. Note that the variations in the composition of the minerals are correlated to the variations in whole-rock chemistry.

Emplacement mechanism of the Tellnes ilmenite deposit (Rogaland, Norway) revealed by a combined magnetic and petrofabric study

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The Tellnes ilmenite deposit (TID) (Sokndal district, Rogaland, Norway) is the second most important ilmenite ore-body in production in the world today (it represented 5% of the world's TiO_2 production in 1999). This world-class Fe-Ti mineralization consists of an ilmenite norite lens-shaped body (> 400 m x 2700 m), that crops out in the central part of the Åna-Sira anorthosite, one of the main geological units of the Late Proterozoic Rogaland Anorthosite Province. The cumulate character of the mineralization is generally accepted (Wilmart et al. 1989, Duchesne & Schiellerup 2001) and the TID has an obvious intrusive character. A time gap of more than 10 Myr separates its emplacement (at 920 ± 3 Ma) from the Åna-Sira anorthosite crystallization (at 932 ± 3 Ma) (U-Pb zircon ages; Schärer et

al. 1996). The TID has not suffered any regional tectonic reworking and metamorphism, in sharp contrast to many North American occurrences of Fe-Ti deposits associated with anorthosites (Duchesne 1999). It thus provides a special opportunity to study genetic and structural primary relationships between anorthosites and ilmenite deposits.

To picture the poorly constrained internal structure of the TID, anisotropy of low-field magnetic susceptibility (AMS) (for a recent review, see Borradaile & Henry 1997) was measured on samples from 49 sites, mainly located in the TID open-pit. In agreement with previous petrographical studies (e.g. Gierth & Krause 1973), the microscopical description of the AMS samples reveals a rather complex magnetic

Fig. 1. (a) Map of the magnetic foliations (planes perpendicular to K3 axes). (b) Map of the magnetic lineations (K1 axes). Schmidt stereo-plots (lower hemispheres) are shown with average values of data (and best axis for Fig. 1a) in the TID and in the Åna-Sira anorthosite (n: number of measurements).

tion (intercept method of Launeau & Robin 1996) performed on about twenty samples further demonstrates that the AMS principal axes are parallel to the corresponding shape principal axes of the opaque subfabric (ilmenite ± magnetite ± sulphides), with an angular departure of usually less than 30°. Since the opaque subfabric mimics the petrofabric of the ore, mainly defined by a common shape-preferred orientation of silicate prismatic crystals (plagioclase + orthopyroxene), this implies that AMS in the TID can be equated with the petrofabric.

The pattern of the magnetic foliations shown in the TID (Fig. 1a) depicts the 3D shape of the orebody, i.e. a SE-plunging elongated trough whose NE and SW flanks dip SW and NE, respectively. This structure presents some apophyses, and ends to the southeast in a dyke-like unit to which is connected a horsetail splay-shaped network of noritic dykes. The pattern of the magnetic lineations (Fig. 1b) is relatively homogenous, from the southeastern network of dykes to the northwestern end of the TID, with a calculated average value at N159° SSE 18°. This orientation is strikingly similar to that of the magnetic foliation best axis (N156° SSE 16°) and is close to the SE-plunging axis of the orebody. Parallelism of the magnetic foliation trajectories with the TID contacts and homogeneity of the magnetic lineation pattern, both observed over the entire orebody, unambiguously point to a fabric that was acquired in a magma flow, during plutonic emplacement of the mineralization. This magma flow can be traced from the magnetic lineation pattern (Fig. 2): it has an average N159° SSE 18° direction, around which some deflections are observed, especially towards some apophyses. The aspect of these deflections, as well as local obliquity between magnetic foliations and the orebody wall, are arguments in favour of a SE to NW (rather than NW to SE) magma lateral spreading along the average N159° SSE 18° direction. Accordingly, the feeding-zone of the TID was located to the southeast, most probably below the horsetail splay-shaped network of dykes connected to the mineralization in that area. The sickle-shaped outcrop of the TID further substantiates that magma injection from the SE feeding-zone was favoured by the transcurrent, dextral opening of a WNW-ESE striking (N120°)

mineralogy. However, simple considerations based on the variation of the bulk magnetic susceptibility (3.1 to 84.3 mSI in the ore) and determination of coercivity spectra suggest that magnetic mineralogy in most samples is dominated by large (usually some tens to several hundreds of mm large) grains of magnetite. Measurements of anisotropy of partial anhysteretic remanence (pAARM) (Jackson et al. 1988) in representative ore samples actually suggest that AMS is mainly carried by these coarse-grained magnetites: pAARM determined after magnetization in a 0-20 mT window, and assumed to be related to the large magnetite grains, is strikingly coaxial with AMS. An image analysis investiga-

weakness zone across the Åna-Sira anorthosite. A previous activation of this weakness zone, at 931 ± 5 Ma (Schärer et al. 1996), would have allowed the emplacement of a several km-long jotunitic dyke (the Tellnes main dyke; Wilmart et al. 1989) to which the TID is associated in a same geological unit.

Two petrogenetic models have previously been proposed for the TID. They take into account the generally admitted cumulate character of the mineralization, and variations in the modal proportion and chemical composition of the ore-forming minerals observed from the contacts towards the central part of the orebody (e.g., decrease of the plagioclase content and increase of the ilmenite one). The first model (Krause et al. 1985, Force 1991) calls for in-situ crystallization of a noritic magma, inward from the orebody flanks. The second model (Wilmart et al. 1989) envisages injection of successive crystal mush batches, coming from the progressive tapping of a magma chamber containing a zoned cumulate pile. The model of Wilmart et al. (1989) is constrained by: (1) the coexistence of strained, prismatic (= cumulus) and undeformed, interstitial (= intercumulus) crystals of plagioclase and orthopyroxene in thin-sections; and (2) the existence of mixing linear trends between ilmenite-rich norites and more plagioclase-rich rocks in variation diagrams of major element contents. Hence, the magma-flow-related fabric we have demonstrated in the TID possibly records the emplacement and the deformation in the sub-magmatic state of successive batches of noritic crystal-mush. This does not preclude, however, that in-situ crystallization has not occurred. On the contrary, the undeformed, interstitial silicates were obviously produced through this process. Moreover, the possibility that the strained, cumulus silicates actually crystallized and were deformed during emplacement in the TID cannot be completely ruled out.

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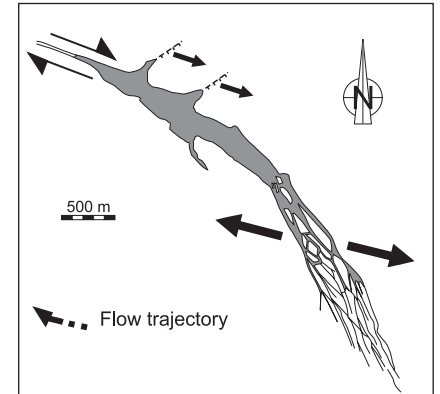
Fig. 2. Tectonic model proposed for emplacement of the TID.

Could the Tellnes ilmenite deposit have been produced by in-situ magma mixing?

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Wilson et al. (1996) argue convincingly that ilmenite-rich cumulates in the 930 Ma Bjerkreim-Sokndal Layered Intrusion, exemplified by rocks near the base of Megacyclic Unit IV (Fig. 1), resulted from mixing of differentiated and more primitive jotunitic magmas. The hybrid was temporarily within the ilmenite phase volume, resulting in precipitation of ilmenite alone, after which co-saturation in plagioclase and orthopyroxene was re-established. Mixing events produced layers 10-20 cm thick with 50% or more modal ilmenite. Such layers that extended across the ~625 km² floor of the magma chamber could contain >31,250,000 m³ of ilmenite.

Modal and cryptic layering in the 920 Ma, dyke-like Tellnes orebody are not obvious. Nevertheless, bulk analyses of thousands of mine samples and cores show that more evolved rocks occur near margins and the most primitive in the interior (Fig. 2). Analyses of selected specimens indicate a range of mineral compositions (An_{37-51}, En_{65-81}, Fo_{71-76}) similar to but exceeding that in the ~90 m-thick regressive zone across the MCU III-IV boundary of the Bjerkreim-Sokndal Intrusion (An_{44-54}, En_{66-76}, Fo_{74}).

Equivalent mineral compositions in both intrusions suggest an origin from similar magmas and mixing events, but within chambers of widely different shape. In Tellnes, mixing occurred in a narrow, steep-sided chamber. Earlier we suggested that ilmenite-rich cumulates formed at the base due to density sorting or in-situ crystallisation along walls as new magma flowed from a conduit at the end or side of the chamber and residual liquid was displaced upward. This called upon jotunitic magma and a process documented in the same district rather than an Fe-Ti-rich magma for which there is no experimental support (Skjerlie et al., Lindsley, this volume).

We have reconsidered the analytical data and our petrogenetic model. The initial model appealed to injection of primitive jotunite magma into a chamber full of differentiated magma. Mixing forced the hybrid well into the ilmenite field, resulting in accumulation of ilmenite on the chamber floor. This would have two consequences: 1) Ilmenite would have a limited composition range, 2) Ilmenite crystals would fall into, trap and react with more evolved magma. We explored ilmenite compositions for evidence of a trapped liquid component, taking account of possible subsolidus effects. The mineral data is inferred to be incompatible with simple mixing (Fig. 3a). Ilmenite, despite scatter from local re-equilibration, shows 1-20% geikielite and correlates with the

Fig. 1. EMP analyses of orthopyroxenes (open symbols) and olivines (closed symbols) from the MCU III-IV boundary at Storeknuten in the Bjerkreim-Sokndal Intrusion. Vertical scale gives stratigraphic thickness in meters. All data from Jensen et al. (1993) and Nielsen et al. (1996). Superimposed are EMP analyses of orthopyroxene from core 7V-3N through the Tellnes orebody from Kullerud (1994). These are plotted using identical height and composition scales as the Storeknuten section, but positioned to emphasize the similar variations.

Tellnes
Section 750 - Bore Hole 7V-3N
(Identical Scales)

Bjerkreim-Sokndal
Storeknuten - III-IV Regression
(Identical Scales)

Fig. 2. Vertical NE-SW section 750 across the Tellnes ore body showing compositions of analyzed orthopyroxenes in cores 7V-3N and 8V-2S. Compositions were contoured by hand according to authors' prejudice.
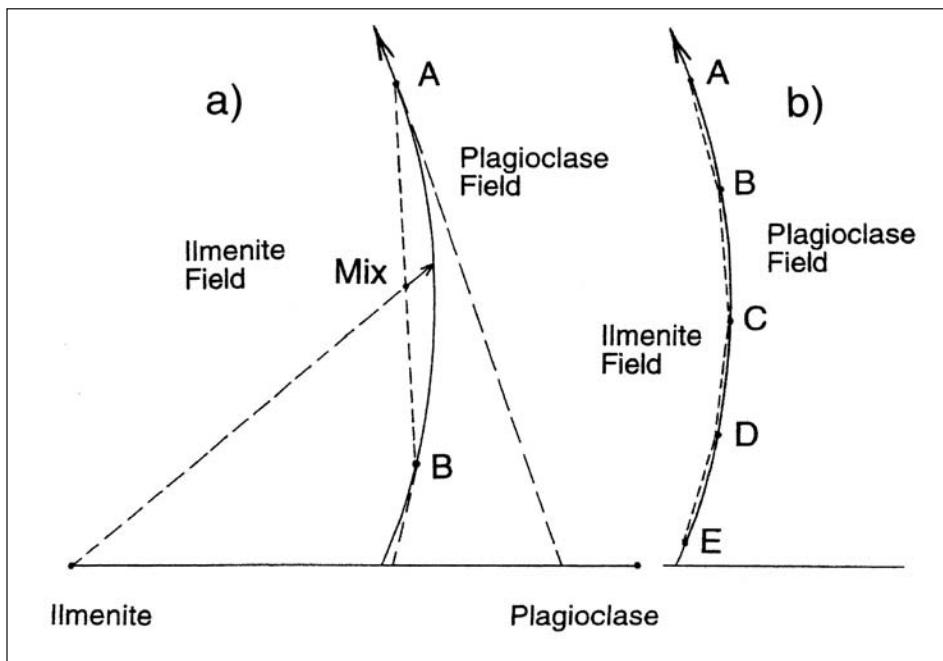
Fig. 3. Diagrams illustrating a hypothetical curved ilmenite-plagioclase cotectic and single a) or multiple b) magma mixing. Due to the curvature (see a), when differentiated cotectic magma A is mixed with primitive cotectic magma B, the hybrid lies entirely within the ilmenite field. Mixed magma, will precipitate only ilmenite, moving directly away from ilmenite composition until reaching the cotectic where plagioclase rejoins ilmenite. A corollary of the curvature is that more primitive magmas (such as B) must precipitate a higher proportion of ilmenite than more evolved magmas (such as A). Instead of a single mixing, as in a), b) illustrates multiple mixing in a dike-like chamber of successive pairs of magma batches that had been progressively evacuated from a chemically, thermally and density stratified subjacent magma chamber. While the net shift into the ilmenite primary crystallisation field is less than in a), there is a still precipitation of excess ilmenite. In addition, upon returns to cotectic precipitation, silicate compositions will reflect melt compositions close to those that precipitated the ilmenite.

most Mg-rich orthopyroxene in each rock ($En_{67-81.5}$). An in plagioclase is also scattered, but the most sodic plagioclases (An_{36-43}) occur with Fe-rich pyroxenes (En_{67-68}), and the most calcic (An_{44-50}) with the most Mg-rich pyroxenes (En_{77-81}). Bulk analyses of the ore average intervals meters thick and may not be comparable to mineral analyses from a single thin-section within the interval. However, Kullerud (1994) showed that abrupt compositional changes are rare in the Tellnes orebody, justifying comparison of compositions of ore intervals and minerals in thin-sections. Due to ilmenite control, there are no meaningful relationships between wt.% TiO_2 vs. other oxides, but wt.% TiO_2 correlates with En in orthopyroxene (7% TiO_2, En_{65-68}, to 25% TiO_2, En_{77-81}). Two intervals with 28 and 33% TiO_2 both contain En_{80-81}. TiO_2 and Mg in ilmenite are correlated (7-14% TiO_2, 1-4% geikielite to 25% TiO_2, 10-19% geikielite). The samples with 28 and 33% TiO_2 contain the most uniform and most Mg-rich ilmenite (18-20% geikielite). TiO_2 correlates also with An in plagioclase (7% TiO_2, An_{37-41} to 25% TiO_2, An_{44-50}). The samples

with 28 and 33% TiO_2 contain An_{46} and $An_{47.5}$, possibly due to a trapped liquid component. In intervals with analyzed minerals, normative An in the bulk ore correlates nearly linearly with En (An_{41}, En_{67}, to An_{54}, En_{81}) and wt.% TiO_2. An interval with normative An_{59} is characterised by En_{81}.

These chemical and mineralogical variations do not support ilmenite crystallisation from a single hybrid magma and evidence for trapped evolved liquid is lacking. There is an alternative way in which mixing may have occurred, but it requires special circumstances and appropriate plumbing. We propose a stably-stratified magma chamber of wide aspect ratio, with differentiated, light magma at the top and more primitive, denser magma at the base (Campbell 1996). The Bjerkreim-Sokndal magma chamber was probably stratified in this manner at times (Wilson et al. 1996). The top of the chamber was tapped and the stratified magma injected into the Tellnes dyke. Adjacent magma layers mixed as they were extracted (Fig. 3b) and the resulting hybrids were saturated in ilmenite alone, although only briefly. Otherwise the magmas were multiply saturated. As the hybrid magmas flowed along the Tellnes chamber, probably from the southeast (Bolle et al., this volume), ilmenite accumulated in higher than cotectic proportions. The density of the resulting cumulates expelled trapped melt effectively. As gradually more primitive magmas from the subjacent chamber were tapped and mixed, the proportion of ilmenite increased. Ilmenite and other phases became more primitive and the central ilmenite-rich zone precipitated from the last and most primitive hybrids. Meanwhile, residual magmas were displaced laterally along the Tellnes dyke, to locations not preserved. Relatively evolved rocks at the top of the Tellnes orebody may have crystallised in situ during injection, though their moderately high ilmenite content is conceptually difficult to explain.

More mineralogical, textural and modal data are required to test the mixing hypothesis and could provide incisive predictive tools for future mining and ore beneficiation at Tellnes.

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Preliminary melting experiments on the Tellnes ilmenite norite from 0.5 to 1.2 Gpa: implications for the composition of intercumulus melt

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There is general agreement that the Tellnes ilmenite norite is a magmatic intrusion that was emplaced at c. 0.5 GPa as a cumulate with minor amounts of intercumulus melt. The composition of the interstitial melt is unknown, but it has been assumed that it was Ti-rich. If the cumulate hypothesis is correct, partial-melting experiments should yield important information on the composition of the melt and equilibrium solid phases. In accordance with this line of reasoning we have performed melting experiments in a piston-cylinder apparatus on a sample of the Tellnes ilmenite norite from 0.5 to 1.2 Gpa and at various temperatures.

The experiments show that the intercumulus melt at 0.5 GPa and T=1050°C (solidus <950°C) is jotunite-like with SiO_2 58.9 wt.%, TiO_2 2.2, Al_2O_3 17.9, FeO 7.2, MnO 0.1, MgO 3.2, CaO 5.3, P_2O_5 1.3, Na_2O 2.3, K_2O 1.6, and in equilibrium with abundant euhedral plagioclase (An_{50}) and ilmenite (He_{15}) together with some olivine ($Fo_{72.5}$). The interstitial glasses in near-solidus experiments are not particularly rich in TiO_2. This is not surprising and is

observed in other melting experiments with residual ilmenite or rutile. The amount of a component like TiO_2 will be buffered by ilmenite or rutile, but the amount of dissolved TiO_2 in a liquid is independent of the amount of the saturated phase. The melt fraction increases with increasing pressure and temperature due to progressive dissolution of plagioclase and ilmenite, and the melt compositions become FeTi-dioritic/FeTi-basaltic. High-temperature melting yields homogeneous glasses with very high abundances of dissolved ilmenite, showing that Ti-rich liquids can exist without the intervention of liquid immiscibility.

Our experiments support the hypothesis that the Tellnes intrusion is an ilmenite-rich, plagioclase-ilmenite cumulate, but the intercumulus melt was jotunite-like and not particularly rich in TiO_2. Increasing dissolution of Fe and Ti at higher P and T suggests that at least some cumulation of ilmenite may have occurred during decompression. Our experiments do not support an origin of the Tellnes ore by liquid immiscibility.

† Deceased 7 september 2003.

Concentration of ilmenite in the late-Sveconorwegian norite-anorthosite Hakefjorden Complex, SW Sweden

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The c. 916 Ma Hakefjorden Complex (HFC) comprises a composite norite-monzonorite-anorthosite intrusion (Åre-bäck 1995, 2001, Scherstén et al. 2000). It is situated c. 25 km NNW of Göteborg, SW Sweden, in the Sveconorwegian Prov-

ince and forms two E-W elongated bodies, both about 500 m wide and 3.6 and 1.2 km long, respectively, intruding Mesoproterozoic supracrustal gneisses (Fig. 1). The intrusion is dominated by a medium-grained norite-monzonorite (An_{54}) contai-
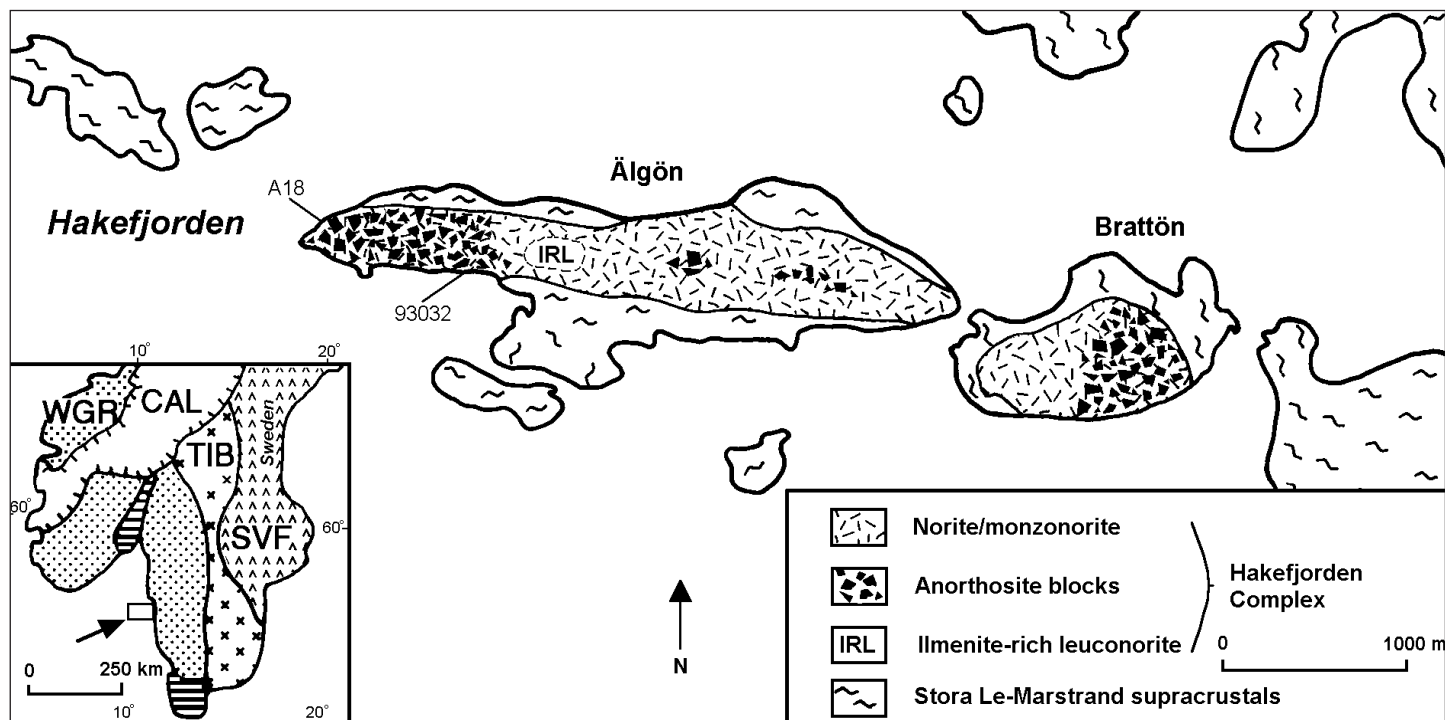
Fig. 1. Generalised bedrock map of Hakefjorden Complex region, SW Sweden. The inset map shows the major lithotectonic units of southwestern Fennoscandia. CAL = Caledonian; SVF = Svecofennian; TIB = Transscandinavian Igneous Belt; WGR = Western Gneiss Region; dotted areas = rocks affected by the Sveconorwegian Orogeny; ruled areas = Phanerozoic rocks.

ning massive anorthosite blocks, 2 cm to 50 m in diameter, and 2-20 cm-large andesine megacrysts (disaggregated anorthosite blocks). The coarse-grained anorthosite blocks (An_{45}) occur in all mafic rock types with a preferred occurrence along the contacts. Subordinate ilmenite-rich leuconorite, hybrid rock and granitic differentiates are recorded. Gradual transitions (mineralogy and geochemistry) between the norite and the marginal monzonorite and the granitic differentiates as well as between the norite and the ilmenite-rich leuconorite, indicate their close relationship and that they are products from the same magma. The emplacement mechanism has been inferred to involve deep-crustal (high-pressure) anorthosite and ultramafic crystallisation followed by mid-crustal norite-monzonorite crystallisation involving crustal contamination and Fe-Ti enrichment during fractional crystallisation (Årebäck & Stigh 1997, 2000).

The liquid composition from which the HFC-norite/monzonorite crystallised is not straightforward to determine, since most rocks are affected by accumulation of cumulus plagioclase ± orthopyroxene ± Fe-Ti oxides ± apatite, crustal contamination or both. However, a slightly more fine-grained margin (chilled margin) where the marginal monzonorite is not developed is suggested to represent the

best estimate of the liquid composition of the exposed HFC-norite/monzonorite (Årebäck 2001). In a QAP-diagram this rock plots in the lower left part of the quartzjotunite (quartzmonzonorite) field. Rocks that are chemically more primitive than this composition are, indicated from thin-section observation, affected by accumulation of ± plagioclase ± orthopyroxene ± Fe-Ti oxides ± apatite, while rocks more evolved are affected by fractionation ± crustal contamination.

In the central part of the intrusion there is a limited area consisting of an extremely ilmenite-rich leuconorite (IRL; Årebäck & Stigh 2000). The IRL occurs as a 'disc-like', c. 10-15 m-thick, 100×300 m-wide subhorizontal unit within the norite. The lateral and lower contacts with the surrounding norite are gradual. The IRL matrix is fine- to medium-grained and characterised by a subhorizontal magmatic foliation defined by the preferred orientation of tabular plagioclase crystals (An_{55}). This foliation is restricted to the IRL unit. Various types of enclaves occur in the IRL matrix. They can be divided into two main groups. The first consists of decimetre- to meter-sized anorthosite and norite blocks from the host HFC. The second group consists of massive to semi-massive, millimetre- to decimetre-sized Fe-Ti oxides, fine-grained anorthosite and fine-grained, fine-scale, igneous-layered plagioclase-orthopyro-
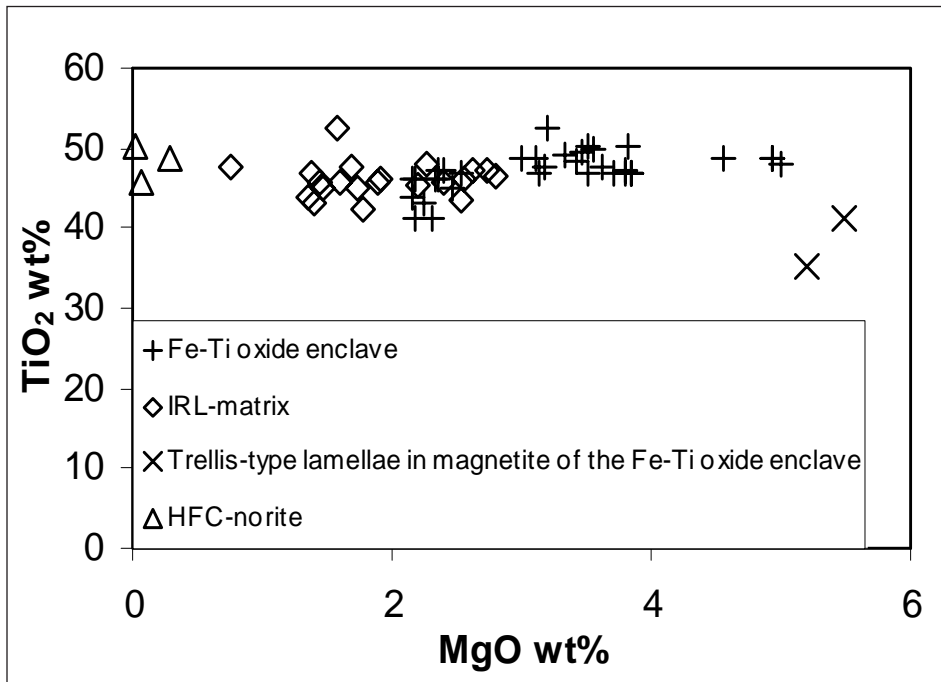
Fig. 2. MgO-TiO$_2$ variation diagram for ilmenites in different units of the HFC. The variation is explained by subsolidus re-equilibration.

xene. The second group is restricted to the IRL and is considered to represent disrupted layering within the IRL. The foliation of the IRL matrix is wrapped around all inclusions and interpreted to result from deformation in the magmatic state.

The IRL matrix contains subequal amounts of ilmenite (32-66 vol. %) and silicates (34-68 vol. %). For obvious reasons, the IRL matrix is anomalously high in major and trace elements entering Fe-Ti oxides, i.e. Ti, Fe, Cr, Sc, Ni and V. Ilmenite grains are typically rounded, equidimensional, and either homogeneous or contain thin hematite lenses arranged in the {0001} planes (hemo-ilmenite). Ilmenite in the massive Fe-Ti oxide enclaves is high in MgO, ranging from 2.17 up to 5.00% (average 3.21%), compared to the IRL-matrix ilmenite (1.35 to 2.79%; average 1.97%), which in turn is more Mg-rich than ilmenite in the surrounding norite (0.03-0.3%; average 0.13%; Fig. 2). The reason for this is interpreted as due to subsolidus re-equilibration; Mg will decrease in ilmenite during cooling by re-equilibration with the surrounding silicates. Oxide-rich units may retain relatively high contents of Mg, since they are more or less isolated from silicates. This explains why the ilmenite in the Fe-Ti oxide enclaves shows higher Mg contents than in the surrounding IRL matrix which, in turn, is more Mg-rich than ilmenite in the norite (Fig. 2). Magnetite rarely exceeds 1 vol. %

in the matrix whereas the Fe-Ti oxide enclaves contain 20-60 vol%. Ilmenite intergrowths occur as microlamellae along the {111} planes (trellis type) of magnetite or as thicker lamellae restricted to one set of the {111} planes (sandwich type). The trellis type is most frequent and occurs in at least two generations; a rather coarse set and a network of very fine-grained lamellae represented by submicroscopic exsolutions occurring as domains in the magnetite. Exsolution of pleonaste is frequent in the {100} planes of magnetite. The lamellae occur in several generations and tend to be most frequent in the centre of the host grain. Textural evidence indicates that a certain amount of pleonaste was exsolved simultaneously with ilmenite, as indicated by the presence of pleonaste microcrystals within exsolved ilmenite. All magnetite has a very low X$_{usp}$ component. However, this does not necessarily mean that the magnetite phase was low in TiO$_2$ during crystallisation, as their compositions are affected by subsolidus re-equilibration.

Crystal accumulation in a late-stage noritic magma residual after anorthosite crystallisation is put forward to be the main mechanism forming the IRL, where the liquid from which it was produced must have been strongly enriched in Fe and Ti. The origin of such liquids is characteristic for residual liquids after anorthosite crystallisation (e.g. McLelland et al. 1994,

Emslie et al. 1994). Deformation in the magmatic state arranged the solid phases according to the stress regime.

Textures and compositions of ilmenite and magnetite in the Fe-Ti oxide enclaves indicate post-crystallisation modification. These enclaves were probably formed, at least in part, by annealing, i.e. consolidation of loose grains into polycrystalline materials at subsolidus temperatures.

Despite the differences in size of the Hakefjorden IRL and the Tellnes ilmenite deposit, similarities in age, petrography and geochemistry occur. The Tellnes ilmenite-rich norite (920 ± 2 Ma; Schärer et al. 1996), which dates the end of the magmatic activity in the Rogaland Anorthosite Complex, is interpreted as a noritic cumulate emplaced as a crystal mush with Fe-Ti-rich intercumulus liquid (e.g. Duchesne 1999).

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Gabbro-hosted ilmenite deposits in Finland

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Ilmenite-rich gabbros in the Kälviä (Kärkkäinen et al. 1997), Kauhajoki (Kärkkäinen & Appelqvist 1999), and Kolari areas (Karvinen et al. 1988), and the Otanmäki Fe-Ti-V mine (closed 1985) (Pääkkönen 1956) indicate a high-Ti potential of small mafic intrusions in Finland (Fig. 1).

Kälviä area, the Koivusaarenneva gabbro

In the Kälviä area, several ilmenite-rich gabbros have been located in a chain of gravity and magnetic anomalies (Kärkkäinen et al., 1997, Sarapää et al. 2001).

The mafic intrusions were emplaced at 1881 Ma into tonalitic bedrock, and belong to a large gabbro province interpreted to have been formed in tensional zones in the vicinity or along the margin of convergent plate boundaries.

The Koivusaarenneva gabbro is an elongated, 0.5 to 1 km thick and 3 km long sill-like intrusion. It can be divided into three different zones, Lower, Middle and Upper, and the main rock type in all zones is a metamorphosed gabbro or gabbronorite (Kärkkäinen 1997). The characteristic minerals in theses zones are ilmenomagnetite, ilmenite and apatite, respectively. The Lower Zone contains minor ilmenite-rich layers with much ilmenomagnetite, and the TiO_2/Fe_2O_3TOT ratio of 0.2 is constant for all rock types in the Lower Zone. The Middle Zone contains a 1.5 km-long and from 50 to 60 m-thick mineralized layer that grades between 8 and 48% ilmenite, 2 to 25 % magnetite and minor ilmenomagnetite. The ratio of ilmenite to magnetite is usually from 3:1 to 4:1. The TiO_2/Fe_2O_3TOT ratio of the Middle Zone is between 0.23 and 0.50. Ilmenite is the dominant Fe-Ti oxide in the Upper Zone that consists of P-rich gabbro and leucogabbro.

The three zones are interpreted to have crystallized from successive pulses of Ti-rich mafic magma. The parent magmas for the Lower and Middle Zones are similar, and the magma for the Upper Zone was more evolved. The Lower and Upper Zones are interpreted to have been generated by relatively closed-system fractional crystallization. The genesis of the Fe-Ti oxide-rich layers in the Middle Zone can be explained by deposition from several magma pulses in an open system, because of the small volume of the rock in the Middle Zone. The Middle Zone may represent a channel for magma flow and the large mass of oxides was accumulated from multiple magma pulses. Suspended oxides were removed from a Ti-saturated magma and sank to the floor of a shallow magma chamber to form the mineral deposit, composed of a Fe-Ti oxide-rich matrix between a silicate framework. The Ti- and Fe- depleted magma flowed out of the poorly crystallized intrusion and was replaced by a new pulse of Ti-saturated parent magma.

Fig. 1. The location of some Finnish gabbro-hosted ilmenite-rich deposits

Kauhajoki area, the Kauhajärvi gabbro

The Kauhajärvi gabbro is one of the several Fe-, Ti-, and P-rich mafic intrusions within a granitoid area in mid-southwest Finland (Kärkkäinen & Appelqvist 1999). It is composed of a relatively thin basal zone and a main zone. Ti- and P-rich gabbro is the most voluminous rock type in the 800 m-thick main zone that is differentiated with a composition varying from peridotite to anorthosite. Layers with an average of 20 wt % combined ilmenite, apatite and ilmenomagnetite form a low-grade apatite-ilmenite-magnetite deposit. Apatite and Fe-Ti oxides are concentrated together with Fe-Mg silicates.

Early crystallization of magnetite supports the interpretation that the main zone crystallized under conditions of relatively high oxygen fugacity. There are three main factors controlling the concentration of the Fe-Ti oxides and apatite in the Kauhajärvi gabbro: 1) an Fe-, Ti- and P-rich parental magma; 2) a relatively high f_{O_2} during crystallization that kept the solubility of Fe and Ti low, so that these elements could not be enriched during crystallization; and 3) high P_2O_5 in the magma enabled the crystallization of individual ilmenite coevally with Ti-bearing magnetite (ilmenomagnetite). The Kauhajärvi intrusion is interpreted to have crystallized under relatively closed-system conditions and the lack of dynamic magma flow through the intrusion prevented accumulation of a large mass of Fe-Ti oxides.

Table 1. Characteristics of some Finnish gabbro-hosted ilmenite-rich deposits. (MT = magnetite, I=ilmenite).

	Koivusaarenneva, Kälviä	Kauhajärvi Kauhajoki	Otanmäki Vuolijoki	Karhujupukka Kolari
Intrusion	multistage gabbro with 3 zones	2-stage well- differentiated gabbro	differentiated gabbro -anorthosite	differentiated sill-like gabbro
Dimensions	0.8 x 3 km^2	>1 x 6 km^2	15 x 1 km^2	0.5 x 1 km^2
Age Ga	1.881 Ga	1.875 Ga	2.06 Ga	2.1 Ga?
Country rock	tonalite	granodiorite	granite-gneiss	metasediments
Geotectonic setting	arc near plate margin	intracratonic rift	shield (continental margin)	cratonic basin
Deposit:				
% TiO$_2$ [1]	7.5	5	14	9
wt-% ilmenite	15 (8 to 48)	10	28	18
wt-% magnetite	6 (-30)	2 - 8	35	66
I / MT	3:1 to 4:1	1.5 : 1	1:2	1: 2 (varies)
% V in MT	0.7	0.3	0.6	0.57
Special		apatite	Co in pyrite	

1) averaged TiO$_2$ content in mineral resource estimates

Summary

Gabbroic intrusions with ilmenite-dominant, Fe-Ti oxide-rich rocks occur within variable geotectonic settings of Palaeoproterozoic ages between 2.1 Ga and 1.88 Ga (Table 1). The common features are the small size of the intrusions and the occurrence of ilmenite and magnetite, mainly as individual grains. Fractionation under low oxygen fugacity conditions favours generation of a Ti-rich parental magma. Shallow magmatic processes concentrated ilmenite from the Ti-enriched parent magma and furthermore, a low f_{O2} environment favoured ilmenite crystallization relative to Ti-magnetite. A magma-flow-through system in the shallow upper crustal environment is the critical process that enables the concentration of large volumes of oxides within small intrusions, as has been suggested for the Kälviä area (Kärkkäinen & Bornhorst 2000). The magma-flow genesis is supported by the fact that the Koivusaarenneva gabbro is a member of a chain of small, probably interconnected, intrusions, three of which, Lylyneva, Peräneva, and Kairineva also, host ilmenite deposits (Sarapää et al. 2001).

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The Grenville Province in Canada

Review of Fe-Ti ± V and Fe-Ti-P$_2$O$_5$ ± V deposits associated with anorthositic suites in the Grenville Province, Québec

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Anorthositic suites are a characteristic feature of the Grenville Province. Numerous deposits of titaniferous magnetite, magnetite-ilmenite-apatite and massive ilmenite have been known since the mid-1850s. However, only a few of them have been economically mined since 1870 and only the world-class, massive ilmenite deposit of the Tio Mine has been mined since 1950. We present a brief review of the most typical Fe-Ti and Fe-Ti-P$_2$O$_5$ deposits associated with anorthositic suites in the Grenville Province in Québec.

In general, the shape of these deposits varies from tabular intrusions to stocks, sills or dykes. Locally, some deposits are characterised by stratiform mineralization in layered anorthositic rocks. Four main distinct styles of mineralization have been described and observed in the Quebec's Grenville Province:

1) Syngenetic formation for disseminated Fe-Ti oxides in the host anorthosite, such as those observed at the west margin of the Havre-Saint-Pierre anorthositic suite.

2) Irregular to concordant injections or intrusions, with clear-cut or diffuse contacts with the host rocks, in an older
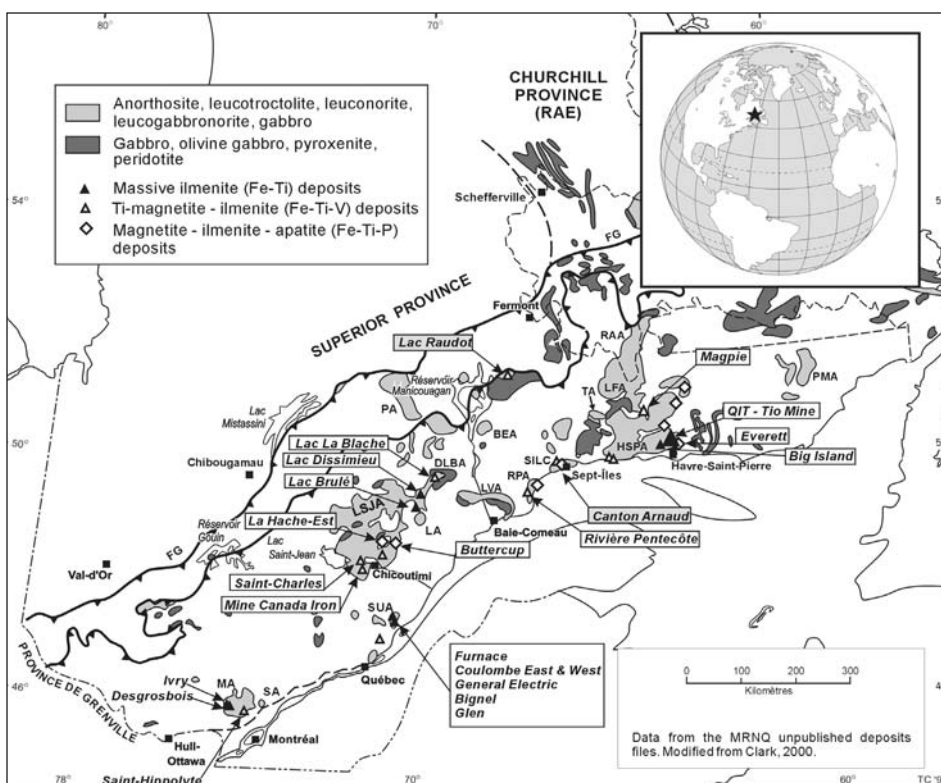
Fig. 1. Location of the studied area in the Grenville Province. HSPA, Havre-Saint-Pierre anorthosite; LFA, lac Fournier anorthosite; LSJA, Lac Saint-Jean Anorthosite; RPA, Rivière Pentecôte Anorthosite; TA, Tortue Anorthosite; PMA, Petit-Mécatina Anorthosite; RAA, Atikonak River Anorthosite; BEA, Berté Anorthosite; LA, Labrie Anorthosite; LVA, Lac Vaillant Anorthosite; PA, Pambrun Anorthosite; MA, Morin Anorthosite; SUA, Saint-Urbain Anorthosite; SILC, Sept-Iles Layered Complex; FG, Grenville Front.

anorthosite or associated rocks, of massive ilmenite, such as the Tio Mine (Fig. 1 and the abstract of Perreault, this volume), the Ivry *(typical chemical analysis for the massive ore is 42.5 % Fe, 33.23 % TiO_2, 7.54 % SiO_2)*, the Bignell *(mineral reserves at 3.44 Mt 34.1 % Fe and 34.7 % TiO_2 with 2.5 Mt at 35.7 % Fe and 36.4 % TiO_2 proven reserves)* and the Lac Brulé deposits (Fig. 1); massive titaniferous magnetite, such as Puyjalon (Everett), Desgrobois and Magpie deposits *(estimation of mineral reserve at 1.8 billions tons with 184 Mt of proven and 627 Mt of probable reserves at 43.1 % Fe, 10.6 % TiO_2, 1.55 % Cr and 0.19 % V)*, oxide-rich norite, ferrodiorite and nelsonite, such as the Puyjalon, Desgrobois, and most deposits in the Lac-Saint-Jean anorthositic suite.

3) Rutile-bearing ilmenitite or oxide-rich norite. Irregular to concordant injections or intrusions, such as the Coulombe East and West *(the mean grade is 36 % Fe, 39 % TiO_2, 4 % SiO_2, 4 % Al_2O_3 and 4 % CaO)* and the Big Island deposit *(typical grab samples collected for assays give values of 26 % FeO, 21.1 % Fe_2O_3, 40.3 % TiO_2, 3.12 % MgO, 0.36 % V_2O_5, 4.35 Al_2O_3, and 6.26 % SiO_2)*.

4) Remobilized Fe-Ti massive or semi-massive mineralizations in shear zones and in late felsic injections cutting the anorthositic rocks, such as those observed at the margin of the Havre-Saint-Pierre anorthositic suite.

A study of mineral compositions of the Lac Mirepoix layered complex, Lac St-Jean Anorthosite Complex (Québec, Canada)

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The Lac Mirepoix layered complex is located within the Lac St-Jean Anorthosite Complex of the Grenville Province, 250 km north of Québec City (Canada). The mineral sequence is very similar to that of the Bjerkreim-Sokndal layered intrusion in Norway. In the sequence of layered rocks, cumulus assemblages contain first plagioclase and hemo-ilmenite, followed by the addition of orthopyroxene, clinopyroxene and magnetite and apatite together as the last phases (Fig. 1). Nelsonite and Fe-Ti oxide mineral pure horizons (10 cm thick) are present in the layered series. Cryptic layering is observed in the composition of plagioclase, orthopyroxene and clinopyroxene. Reversals in the mineralogical sequence and in the evolution of the An content of plagioclases and Mg# of orthopyroxenes and clinopyroxenes are interpreted as due to a new magma influx between the Cyclic Units IV and V (Fig. 1). Mixing of the new magma influx with the resident magma explains the intermediate mineral compositions at the base of the new influx (Cyclic Unit V, Fig. 1).

Magnetite and hemo-ilmenite of the Lac Mirepoix layered complex have been affected by strong subsolidus re-equilibration. Coronas of spinel pleonaste are commonly observed at the contact of the two oxide minerals. Analyses of whole grain oxide minerals show that Al_2O_3 and MgO of the ilmenite vary in the same way. These analyses also show that the proportion of hematite in the ilmenite is usually less important when magnetite is present (Fig. 1). This is explained by a two-step process. The first step is the exsolution of a spinel pleonaste phase in the magnetite at the contact of the ilmenite. The exsolutions are incorporated, in a second step, to the ilmenite grain by the formation of new ilmenite at the contact of the oxide minerals grains. This new ilmenite is formed by

oxidation of the ulvöspinel in the magnetite. The oxidation is caused by reaction with the hematite contained in the ilmenite (Buddington & Lindsley, 1964, Duchesne, 1972). Because of this process, Al_2O_3 and MgO from the magnetite are incorporated in the ilmenite. It also lowers the TiO_2 in the magnetite and the hematite content of the ilmenite.

In the hemo-ilmenite of the Lac Mirepoix layered complex, an ulvöspinel-magnetite$_{ss}$ phase is observed as 'exsolution' lamellae. These 'exsolutions' are probably produced by reduction of the hematite exso-

lution lenses. This reduction in the hemo-ilmenite can be coupled with oxidation in the sulfide minerals.

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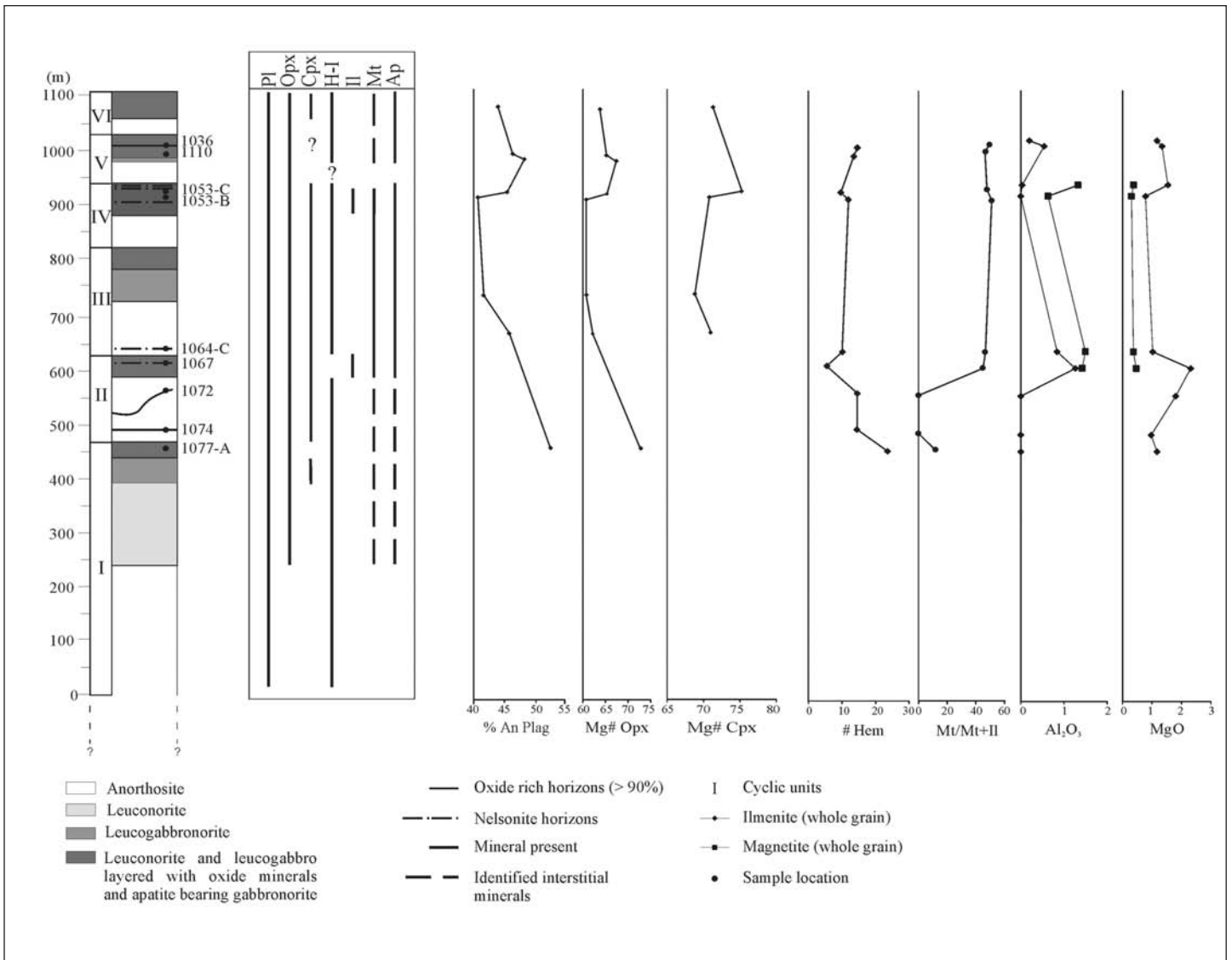
Fig. 1. Generalized stratigraphy of the Lac Mirepoix layered complex, cryptic layering and whole grain oxide compositions. (Pl) plagioclase; (Opx) orthopyroxene; (Cpx) clinopyroxene; (H-I) hemo-ilmenite; (Il) ilmenite; (Mt) magnetite; (Ap) apatite; (% An plag) 100*An/(An+Ab); (Mg# Opx) Mg# of orthopyroxene; (Mg# Cpx) Mg# clinopyroxene; (# Hem) $0{,}5*(Fe^{2+} + Fe^{3+} - Ti + Mg + Mn)$; (Mt/Mt+Il) ratio magnetite on magnetite plus ilmenite; (Al_2O_3, MgO) whole grain oxide minerals by XRF.

The 974 Ma Vieux-Fort Anorthosite, Lower North Shore, Québec: the youngest anorthosite in the Grenville Province

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The Vieux-Fort Anorthosite is a small complex (~ 60 km^2) located in the northeastern Grenville Province, in Quebec's Lower North Shore area (see Fig. 1 of Perreault & Hébert, this volume). The complex is composed mostly of massive to weakly deformed anorthosite and leuconorite with small amounts of leucogabbro, pegmatitic gabbro, and gabbro; it is bordered to the north by mangeritic rocks (Fig. 1). South of the NE-SW-trending Vieux-Fort fault, leuconorite at the contact of the pluton is foliated parallel to the regional foliation in the adjacent Pinware Terrane granitic gneiss.

Zircons recovered from a massive, pegmatitic, quartz-bearing, granophyric leuconorite are colourless to light tan, irregular fragments and shards, with a low uranium content (38-267 ppm U, usually <100 ppm U) and a high Th/U ratio (0.8-0.9),

features typical of zircons crystallizing rapidly at a late stage from a mafic magma. Many grains have tiny, opaque or transparent mineral inclusions; a few grains are fractured. The U-Pb results show a slight scatter but the ^{207}Pb/^{206}Pb ages for all six analyses are clustered between 960 and 976 Ma; they are 0.6-0.9 % discordant except for two analyses at −1.1 % and 1.5 %. A best-fit regression yields an upper intercept age of 974.5 ±1.8 Ma, which is interpreted as a crystallization age (Fig. 1). Thus, the Vieux-Fort Anorthosite is the youngest known anorthosite in the Grenville; younger than the 1018 Ma Labrieville Anorthosite and the 1062 Ma Havre-Saint-Pierre Anorthosite.

The massive appearance of the anorthosite and of the related mangerite envelope, with tectonic foliation limited to the margin of the pluton, and the absence of metamorphic mineral assemblages in the anorthositic rocks, suggest that the Vieux-Fort Anorthosite has escaped the major Grenvillian (1080-980 Ma) deformation and associated metamorphism that affected the eastern Grenville Province. The crystallization of the Vieux-Fort Anorthosite at 974 Ma was coeval with late- to post-Grenvillian gabbros, syenites, monzonites and granites (980-966 Ma) in the eastern Grenville. Pb loss at 950 Ma, suggested by a slight scatter in the distribution of the analyses, is coeval with a cluster of titanite U-Pb ages (970-930 Ma), which correspond to titanite U-Pb closure temperatures and to post-peak metamorphic cooling. We suggest that the Vieux-Fort Anorthosite was emplaced after the main Grenvillian deformation and metamorphism, in a partially cooled, thickened crust. The Vieux-Fort Anorthosite and its associated mangeritic rocks may be coeval with the earliest components of post-tectonic gabbroic and granitic magmatism resulting from melting at the base of the thickened Grenvillian crust.

Fig.1. Simplified geological map of the Vieux Fort area and U/Pb diagram.
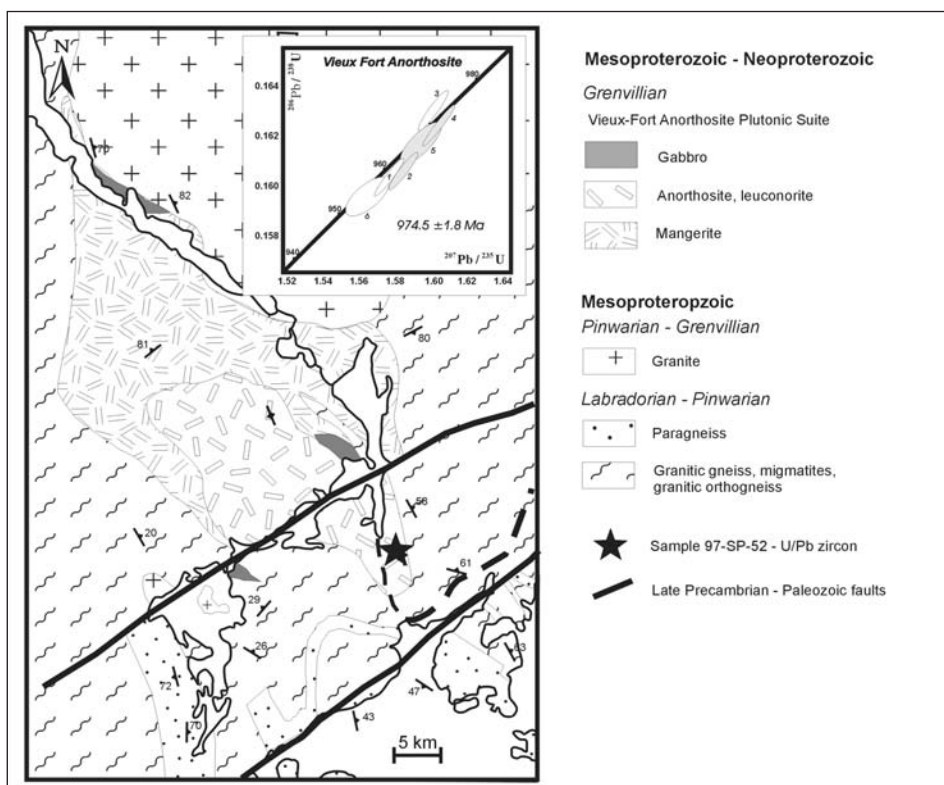
Contrasting styles of Fe-Ti mineralization in the Havre-Saint-Pierre anorthosite suite, Quebec's North Shore, Canada

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The Havre-Saint-Pierre anorthosite suite (HSP) is located in the northeastern part of the Grenville Province (see Fig. 1 of Perreault & Hébert, this volume). It is composed of several massifs or lobes separated from each other by structural zones. It includes a wide range of rock units typical of AMCG suites. The HSP is composed of seven massifs: the well-known Lac Allard massif, as well as the Rivière Romaine, the Rivière Magpie-Ouest, the Nord-Ouest, the Sheldrake, the Brézel, and the Rivière-au-Tonnerre massifs. Most of the Fe-Ti mineralizations are associated with the lac Allard, the Rivière-au-Tonnerre and the Rivière Romaine massifs. The Nord-Ouest massif is characterised by Cu-Ni mineralizations associated with pyroxenite and melanorite in anorthositic rocks at the margin of the massif (Clatk 2000). The structural state of the HSP ranges from undeformed and weakly recrystallized anorthosite to highly deformed and recrystallized anorthositic gneiss. Partially or totally deformed and recrystallized mangerite and charnockite envelop most of the anorthositic massifs. Slivers of deformed and highly metamorphosed supracrustal rocks are commonly found between the massifs. The emplacement of the HSP was complex: in the north, it was thrust over the country rocks; in places, one massif was thrust over another; and in the south, the relatively young Rivière-au-Tonnerre massif was intruded into an older massif. U-Pb zircon geochronology on various massifs shows a range in ages. A mangerite associated with the Allard Lake massif was dated at 1126 ± 6 Ma (Emslie & Hunt 1990). An age of 1130 Ma was found for the Sheldrake massif (in Verpaelst et al. in prep.), and an age of 1062 Ma was obtained for the Rivière-au-Tonnerre massif (van Breemen & Higgins 1993).

In the Havre-Saint-Pierre anorthositic suite, two contrasting styles of Fe-Ti mineralization are present. In the eastern part of the HSP, massive hemo-ilmenite and Fe-Ti-P_2O_5 mineralizations are commonly found. Since 1950, QIT has extracted an estimated 60 Mt of ore grading 38.8 % Fe and 33.6 % TiO_2 and the mean composition of the ore is 34.2 % TiO_2, 27.5 % FeO, 25.2 % Fe_2O_3, 4.3 % SiO_2, 3.5 % Al_2O_3, 3.1 % MgO, 0.9 % CaO, 0.1 % Cr_2O_3, and 0.41 % V_2O_5. Massive hemo-ilmenite deposits are interpreted as immiscible Fe-Ti-rich liquid derived from a parental jotunitic or oxide-rich noritic magma, which intruded a partially crystallized andesine-bearing anorthosite. On the opposite side of the HSP, on the southwest border, Fe-Ti mineralizations are mostly composed of titaniferous magnetite and a lesser amount of ilmenite. The style of emplacement of the mineralization ranges from primary massive magnetite layers in a labradorite-bearing anorthosite to weakly discordant magnetitite and oxide-rich leuconorite veins emplaced roughly parallel to the regional tectonic fabric in the anorthosite. Finally, remobilized magnetite veins, magnetite-rich monzonitic pegmatites and magnetite veins emplaced in late ductile to brittle shear zones cut the HSP anorthosite.

Massive ilmenite dykes and veins

The world-class Tio Mine deposit, exploited since 1950 by QIT, is the classic example of this type. The Tio Mine is located 42 km north of Havre-Saint-Pierre (Fig.1) and was discovered in 1946 by geologists of Kennecott Copper Inc. Over 20 significant showings were discovered by J. Retty in 1941 and by Kennecott and QIT geologists between 1944 and 1950 near lakes Allard and Puyjalon.

The Tio Mine (site 1, Fig.1) and most satellite deposits are located on the east side of the HSP. There, the plagioclase in the anorthosite is andesine and most

LEGEND

Ordovician
Dolomite, limestone, sandstone

Neoproterozoic
Granite

Mesoproterozoic
Gabbro, amphibolite

Anorthosite, leuconorite, norite, gabbro

Mangerite, monzonite, charnockite

Granite

Porphyritic granite and granitic augen orthogneiss

Monzonite

Metagabbronorite, monzonite and porphyritic granite

Granite, foliated and gneissic granite

Meta-volcanosedimentary rocks and metagabbro of the Wakeham Group

Gneissic complex

Undifferentiated gneiss

▼ Massive hemo-ilmenite mineralizations

▽ Ilmenite QIT's Tio Mine

◆ Fe-Ti-P2O5 mineralizations associated with norite / ferrodiorite

▲ Massive titaniferous magnetite mineralizations

■ Ilmenite mineralization associated with oxide-rich norite.

Fig.1. Simplified geological map with the major Fe-Ti showings and deposits of the Havre-Saint-Pierre anorthosite suite (HSP). LAS, lac Fournier anorthosite suite; LAm, lac Allard massif; NQm, Nord-Quest massif; RATm, Rivière-au-Tonnerre massif; RRm, Rivière Romaine massif; Sm, Sheldrake massif; RMOm, Rivière Magpie-Quest massif; Bm, Brézel massif.

deposits are closely associated with jotunite or oxide-rich norite (Rose 1969, Bergeron 1986, Force 1991). These deposits occur as dykes, sills, or stratoid lenses. Most associated norite or jotunite intrusions show an ilmenite enrichment at the base. Jotunite intrusions are composed of plagioclase (An41-44), orthopyroxene, apatite, ilmenite, and magnetite. Locally, modal proportions of apatite and oxides reach 10 % and 20 %, respectively. Recent models suggest that these massive ilmenite deposits originated from the separation of an immiscible iron- and titanium-rich liquid from a jotunitic parent liquid (Force 1991). This immiscible oxide-rich magma then moved into fractures and shear zones in the crystallizing anorthosite to form dykes, sills, and tabular intrusions of massive ilmenite, such as those observed at the Tio Mine (Bergeron 1986, Force 1991).

The Tio Mine is composed of a subhori-zontal, massive ilmenite intrusion forming three deposits separated by normal faults. The main deposit measures 1097 m in the N-S trend and 1036 m E-W. Its thickness is estimated at 110 m and the deposit is inclined 10° to the east. The deposit is divided into 3 zones. The upper zone is composed of massive hemo-ilmenite, the intermediate zone of alternating layers of massive to semi-massive hemo-ilmenite and anorthosite, and the lower zone of massive hemo-ilmenite. The Northwest deposit forms a band 7 to 60 m in thickness of massive ilmenite alternating with anorthosite. The deposit is gently inclined to the east. The Cliff deposit forms a hill that overlooks the Tio Mine, and has an ellipsoidal shape. Both the Main and the Northwest deposits are mined. The ore is composed of a dense, coarse-grained aggregate of hemo-ilmenite. Accessory minerals are plagioclase, spinel, pyrite, and chalcopyrite. Locally, the modal proportion of sulphides reaches 2

%. The Main deposit is estimated at 125 millions tons. The Northwest deposit contains 5 Mt at 37.4 % Fe and 32.32 % TiO_2, and the Cliff deposit contains 8.4 Mt at 39.2 %Fe and 33.9 % TiO_2.

Other similar deposits occur around Lakes Allard and Puyjalon. The most promising deposits include Grader (which was mined in 1950 for a short period), Springer, and Lac-au-Vent. All these deposits and smaller showings are composed of massive ilmenite dykes emplaced in the andesine-bearing anorthosite. One exception worthy of mention is the Big Island deposit (site 2; Fig.1). This deposit is composed of a rutile- and sapphirine-bearing, massive hemo-ilmenite dyke with fragments of oxide-rich norite and anorthosite. Rutile can be as high as 15 modal %. Typical grab samples collected for assays gave values of 26 % FeO, 21.1 % Fe_2O_3, 40.3 % TiO_2, 3.12 % MgO, 0.36 % V_2O_5, 4.35 Al_2O_3, and 6.26 % SiO_2 for a sapphirine- and rutile-bearing ilmenitite, and 29.1 % FeO, 29.8 % Fe_2O_3, 38.9 % TiO_2, 2.99 % MgO, 0.39 % V_2O_5, 1.54 % Al_2O_3, and 0.91 % SiO_2 for massive ilmenitite. The dyke is 15 m wide and crops out over 500 m.

Magnetite-ilmenite-apatite mineralizations

Most ilmenite-magnetite-apatite mineralizations are associated with jotunite on the east side of the HSP (Fig. 1). These rocks are present at the contact between the anorthosite and mangeritic rocks. The jotunite is usually coarse-grained, dense, and highly magnetic. Close to the contact with the anorthosite, a primary layering is well developed (Rose 1969, Hocq 1982, Bergeron 1986). Several deposits were discovered along with the hemo-ilmenite deposits. Two typical deposits are described here. The Everett deposit (Fig. 1) crops out along the northwest shore of Lake Puyjalon. The ilmenite-magnetite-apatite mineralization is associated with a band of jotunite 3 km long and up to 300 m thick. The mineralization is composed of 20 to 30 % ilmenite, hematite, magnetite, and apatite, along with plagioclase and orthopyroxene. Locally, oxides and apatite make up to 50 % of the rock. Gulf Titanium Limited has estimated geological reserves in the deposit at 293 Mt grading 16.2 % Fe, 9.75 % TiO_2, and 4 % P_2O_5.

Other showings are located northeast of the Everett deposit and about 60 km northeast of Tio Mine (Fig. 1). The mineralizations are similar to the Everett deposit and are associated with jotunite emplaced at the contact of the anorthosite with the mangerite. The main showing outcrops over 1 km and is 100 m wide. The mineralization occurs as magnetite- and ilmenite-rich layers from a few centimetres to half a metre in thickness in the jotunite. The rock shows a well-developed banded structure. The jotunite is composed of 50 to 60 % magnetite and ilmenite, 10 to 30 % plagioclase, 10 to 30 % orthopyroxene, and accessory apatite and green spinel.

Massive ilmenite veins with ilmenite-rich leuconorite

This type of mineralization is observed on the southwest side of the HSP (East of Rivière-au-Tonnerre, Fig.1). It is closely associated with leuconoritic rocks intruding the labradorite-bearing anorthosite. The leuconorite forms thick dykes that invaded a partially crystallized anorthosite. The dykes are emplaced parallel to the foliation observed in the anorthosite and contain large enclaves of the host rocks. Contacts between the two rock types are usually sharp but are locally diffuse. Ilmenite (up to 10 %) and orthopyroxene are disseminated in the leuconorite and large corroded and broken xenocrysts of labradorite are common. The labradorite xenocrysts are fragments of the host anorthosite. Massive ilmenite lenses and irregular veins are common within accumulations of orthopyroxene phenocrysts in the norite. The thickness of these lenses and veins varies from 10 cm to 60 cm. The ilmenite veins are intrusive in the leuconorite and locally they cross the contact between the anorthosite and the norite and form thin veins of massive ilmenite in the anorthosite. The ilmenite-rich veins are composed of lamellar ilmenite (60 to 80 %), plagioclase, orthopyroxene, green spinel and locally apatite, clinopyroxene and magnetite. Late biotite after ilmenite is commonly observed. Values of 46.3 to 59.3 % Fe_2O_3, 30.6 to 38.7 % TiO_2, 2.06 to 2.42 % MgO, 0.33 to 0.43 % V_2O_5, 1.64 to 13.0 % SiO_2, 1.35 to 5.90 % Al_2O_3, 0.2 to 1.35 % CaO, and 0.01 to 1.1 % S were obtained on 3 samples of ilmenitite veins.

Titaniferous magnetite mineralizations

Non-economic, titaniferous magnetite mineralizations are common in labradorite-bearing anorthosite and are mostly observed in the southwest part of the HSP (West of Rivière-au-Tonerre, Fig. 1). The mineralizations occur in several forms: (1) layers of magnetitite in anorthosite; (2) layers, discordant veins, or dykes of magnetite-rich melanorite or melagabbronorite; (3) discordant veins and lenses of remobilized magnetitite; (4) magnetite-rich monzonitic pegmatite; and (5) magnetite-rich lenses or veinlets in late ductile and brittle shear zones cutting anorthosite.

Two localities are worth mentioning. At the mouth of Rivière Chaloupe and at Cap Rond, the magnetite mineralization forms centimetre-thick layers of semi-massive and massive magnetite in the anorthosite. Most of the layers are interpreted as primary; however, some of the magnetite-rich layers are associated with the development of a tectonic foliation in ductile shear zones cutting anorthosite. This is particularly well developed at the western margin of the HSP in the Rivière-au-Tonnerre lobe. Locally, in well-exposed outcrops along the shoreline, the layers are folded and locally reoriented in the ductile shear zones. Near the contact with the magnetitite layers, the anorthosite shows an increase in magnetite content. In the same outcrop, remobilized magnetite veins originating from the layers cross-cut the tectonic foliation at a low angle. Fragments of undeformed labradorite xenocrysts are common in these veins. Plagioclase in the host anorthosite is highly deformed and recrystallized.

Veins of magnetitite or magnetite-rich melanorite from a few centimetres to 15 metres thick are quite common in this part of the HSP. These veins cut at a low angle the tectonic and primary foliation and are oriented more or less parallel to the regional tectonic fabric. Large xenocrysts of labradorite and enclaves of the host anorthosite are common. Layered anorthosite enclaves also occur in these magnetite-rich veins. The magnetite content varies from over 70 % in the oxide-rich veins to 30 to 50 % in the melanorite. Other minerals included plagioclase, orthopyroxene, hornblende, green spinel, ilmenite, pyrite, chalcopyrite, clinopyroxene, and apatite. Values of 49.5 % Fe and 16.34 % TiO_2 were obtained on one magnetite-rich vein with a thickness of 15 m. Analyses of thinner veins show that the Fe_2O_3 content varies from 50 % to 66 %, TiO_2 from 10 % to 19.2 %, and V_2O_5 from 0.24 to 0.36 %. However, due to the small scale of these veins and also to the nature of the titaniferous magnetite, they are non-economic.

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Cambrian titaniferous paleoplacers metamorphosed during the Taconian Orogeny, Quebec Appalachians

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Force (1991) stated that only two types of ore deposit account for more than 90% of the world production of titanium minerals: 1) magmatic ilmenite deposits and (2) young shoreline placer deposits. He further noticed that the placer resources are declining while the consumption of titanium minerals is increasing. Force (1991) concluded that "by the year 2010, we are likely to see titanium minerals being produced from deposit types not currently exploited. This means a challenge for economic geologists to define these new ore deposit types". The purpose of this contribution is to address this challenge from a northeastern North American perspective. Regional metallogeny concepts are first applied, e.g., 1 (nature and extent of the proper source rocks, 2) weathering history actual (Quaternary) and past (Eocambrian). Then, a specific location where these favourable conditions seem to have occurred (e.g. Sutton, Québec) will be examined in further detail.

The Grenville Province is a Mesoproterozoic orogenic belt flanking the southeastern margin of the Canadian Shield. It

Fig. 1. Geological sketch map of the Grenville Province and Quebec Appalachians.
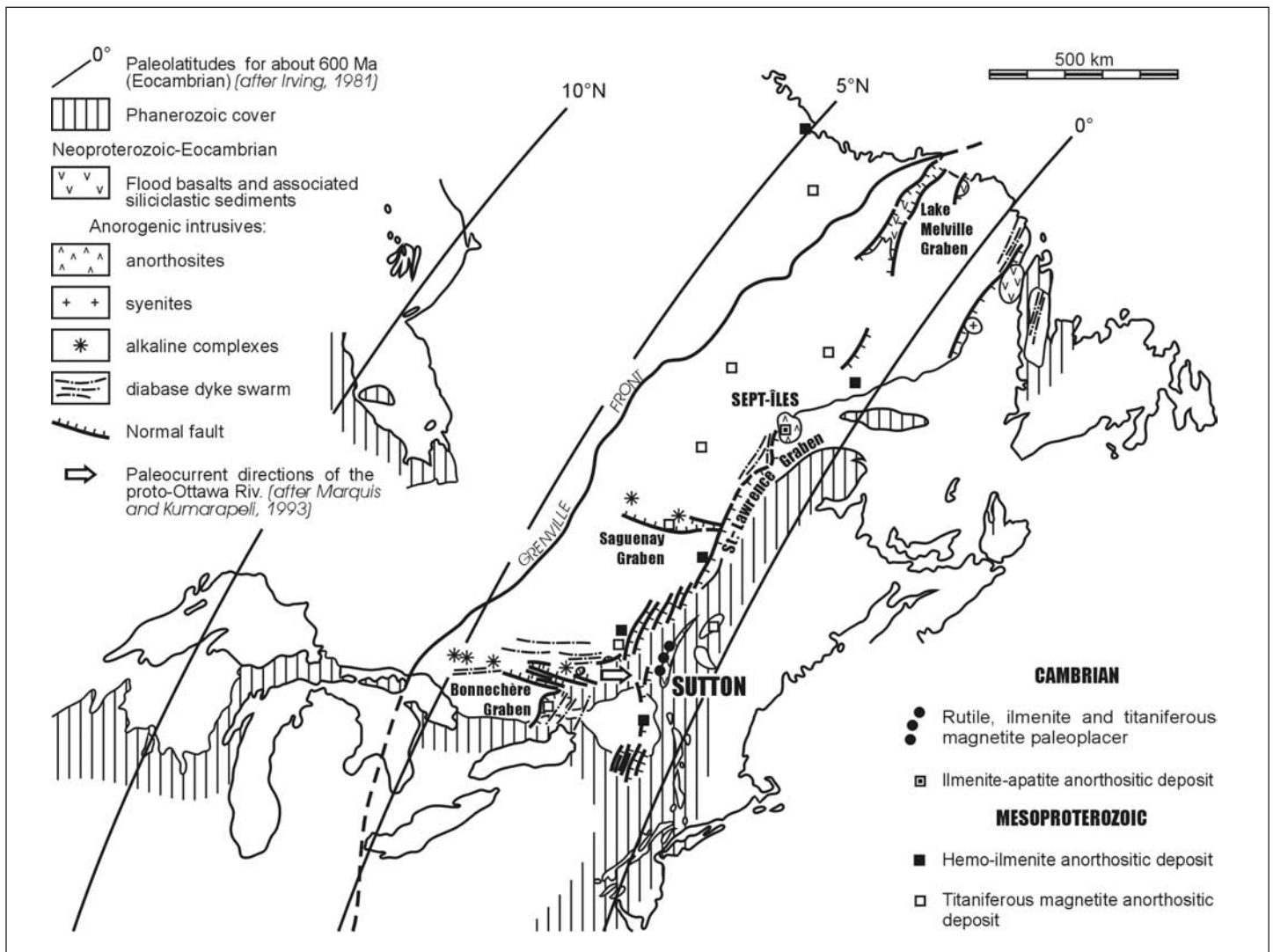
occurs as a 500 km wide and 2 000 km long high-grade metamorphic belt with outliers in the Paleozoic Appalachian orogenic belt of New Jersey, Vermont, Maine and Quebec. The Grenville Province is known for its large anorthosite massifs and for its numerous magmatic titaniferous deposits. The most important of which are Lac Allard in Québec and Sanford Lake in the Adirondack Mountains of New York. Considering all this, the Grenville Province appears to be a very prospective source region for younger titaniferous placers. Indeed, placers are known to occur in Quaternary alluvial and deltaic deposits. The most important is Natashquan, a large deposit occurring in the Natashquan river delta on the St. Lawrence North Shore. Another Quaternary placer occurs where the Peribonka river forms a delta in lake St. Jean. Both the Natashquan and the Peribonka placer deposits have been evaluated in recent years (e.g., Natashquan by Tiomin in the 1990s and Peribonka by SOQUEM in the 1970s). Both deposits are large but do not seem to be economic at the present time.

Proper weathering of primary titanium minerals seems to be a prerequisite for the formation of economic placer deposits. Low-latitude (below 30°) weathering is especially important in increasing the percentage of TiO_2 in the ilmenite concentrate (Force 1991). The duration of weathering under favourable paleolatitude is another important factor. Australia is quite favoured from that perspective. On the contrary, Natashquan and Peribonka, as well as the other Quaternary placers deposited in the Grenvillian region, lack these two prerequisites (i.e. the main weathering process involved was a continental glaciation and the deposition occurred above 45°N).

In the search for new titanium mineral resources, the next question is: if modern conditions were unfavourable for the formation of economic placer deposits in the Grenvillian region, is there any geologic period in which the conditions were more favourable? Such an approach has been successfully applied to other ore deposit types (e.g. stratiform copper deposits, cf. Kirkham 1989).

In Eocambrian (Late Neoproterozoic) time, the Grenvillian cratonic region expe-

rienced deep weathering under a low-latitude climate (Fig. 1). The widespread occurrence of red-bed sediments in rift facies of the early Iapetus Ocean testifies to such a paleoenvironment in the Quebec Appalachians (Gauthier et al. 1994). If paleoplacers of this age are preserved in Cambrian strata, they might be of better grade than their Quaternary equivalents. During the 1980s, such paleoplacers have been identified near Sutton (Fig. 1) in the Quebec Appalachians.

The Sutton paleoplacers occur at the base of the Pinnacle Formation, in wacke and sandstone overlying alkaline volcanic rocks of the Tibbit Hill Formation. These heavy mineral deposits are marine placers that were deposited on an extensive shoal with strong drift currents at the mouth of the Proto-Ottawa river (Marquis & Kumarapeli 1993). Extending laterally for several kilometres, the horizon comprises titaniferous magnetite and ilmenite, replaced by a hematite-leucoxene assemblage, as well as primary rutile, zircon and tourmaline as heavy minerals. Massive heavy-mineral horizons average 3 m and locally are 7 m thick. Together, the opaque minerals comprise 10 to 50 %, and locally 95 %, of the black sand paleoplacers.

The paleoplacer horizons were folded during the Mid Ordovician Taconian orogeny. Detrital titanium minerals were sheared and dispersed along the foliation planes and formed fine-grained leucoxene disseminations. These leucoxene grains form fine intergrowths with chlorite. Local recrystallization of the detrital hematite-leucoxene assemblage to low-Ti magnetite took place along Taconian thrust faults, producing an assemblage of euhedral magnetite and rutile with rounded detrital grains of zircon.

The rutile potential of these paleoplacers was evaluated by SOQUEM in the early 1990s. Although tonnages and grades were impressive, the fine grain size of the rutile and its intimate intermixing with hematite renders commercial extraction of the rutile uneconomic. Although a major source region and favourable climate prevailed during the deposition of these paleoplacers, subsequent deformation under greenschist-facies conditions proved to be deleterious. However, recrystallization along Taconian thrust faults was benefici-

al for rutile recovery.

The Sutton case history suggests that the search for new titanium mineral resources may need the interplay of different disciplines of the Earth sciences (e.g., igneous petrology, paleo-climatology, sedimentology, metamorphic petrology and structural geology), a combination which is not usual in economic geology. The deleterious effects of low-grade metamophism and the positive effect of higher grade recrystallization on the Sutton titanium mineral deposits is reminiscent of the Lake Superior type of iron deposits in which the iron-ore industry make a breakthrough in the 1950s by combining a metamorphic petrology approach with sedimentology (James 1954, 1955). This may be one of the ways of addressing the challenge to economic geologists pointed out by Force (1991).

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Areas in eastern Europe

Ilmenite deposits and mineralization in alkaline and subalkaline magmatic complexes of the Ukrainian Shield

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Several ilmenite deposits related to alkaline and subalkaline complexes are known in the Ukrainian Shield. They usually contain variable contents of apatite, ilmenite and magnetite. Proterozoic alkaline and subalkaline complexes with ilmenite mineralizations predominate. A Devonian alkaline complex (Pokrovo-Kyreyevo) also occurs in the border zone with the Donbass folded structure (Fig.1).

Ilmenite-bearing Proterozoic alkaline and subalkaline complexes are subdivided into two types : alkaline-ultrabasic (with carbo-

natites) and gabbro-syenitic. These types have different geological ages: about 2.1 Ga for the first and 1.7-1.8 Ga for the second.

Ore concentrations of ilmenite associated with the alkaline-ultrabasic complexes are found only in the Chernigovka (Novo-Poltavka) massif situated in the West Azov area. Ore-bearing rocks are represented there by alkaline pyroxenites (nepheline-free jacupirangites), which consist mainly of alkaline pyroxene (aegirine-diopside, aegirine-salite), ilmenite and magnetite, and subordinate amphibole, biotite, apati-
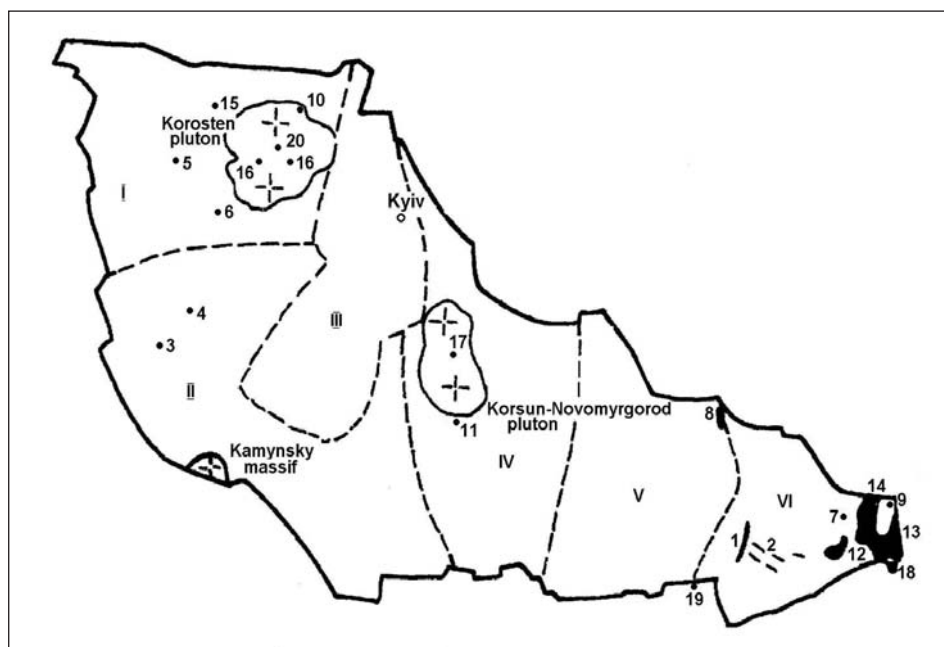
Fig. 1. Massifs and occurrences of carbonatites, alkaline and subalkaline rocks in the Ukrainian Shield. Ultrabasic-alkaline complexes: 1 – Chernigovka (Novo-Poltavka), 2 – dykes of metajacupirangites, 3 – Proskurovka, 4 – Antonovka, 5 – Gorodnitsa intrusive body, 6 – fenites of Berezova Gat. Gabbro-syenitic complexes: 7 – Oktyabrsky, 8 – Mala Tersa, 9 – Pokrovo-Kyreyevo, 10 – Davydky, 11 – Velika Vyska, 12 – South Kalchyk, 13 – Yelanchyk, 14 – Kalmius, 15 – Yastrubetsky, 16, 17 – aegirine syenites of Korosten and Korsun-Novomyrgorod plutons, 18 – Prymorsky, 19 – Melitopol, 20 – Stremygorod apatite-ilmenite deposit. Geological blocks of the Ukrainian Shield: I – Northwestern; II – Dniepr-Bug; III – Ros-Tikich; IV – Ingulo-Ingulets; V – Middle-Dniepr; VI – Azov.

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	S	H_2O-	LOI	CO_2	Total
1	36.90	6.46	2.13	12.63	11.23	0.55	10.61	15.60	0.95	0.60	tr.	0.27	0.09	1.38	0.51	99.98
2	36.40	9.00	2.63	9.80	9.50	0.16	7.87	17.30	1.20	1.60	2.29	0.50	0.15	0.78	0.85	100.03
3	38.54	5.63	3.16	9.92	10.14	0.41	9.84	15.62	1.84	1.05	0.89	0.47	0.13	0.96	1.54	100.14
4	40.32	7.02	12.55	3.01	13.16	0.29	6.82	11.32	2.74	0.40	0.09	0.26	0.08	1.11	0.48	99.65
5	35.60	6.40	12.90	12.55	10.90	0.20	7.30	11.28	1.45	0.35	0.05	0.06	-	-	-	99.04
6	33.85	10.02	2.52	11.12	13.68	0.19	10.50	14.27	0.64	0.40	0.90	0.16	0.18	1.02	0.35	99.80
7	38.86	9.05	3.58	6.56	12.97	0.09	11.72	14.23	0.74	0.39	0.21	0.08	-	1.28	-	99.76
8	31.50	7.22	3.06	5.90	28.60	0.43	6.40	10.86	0.72	0.45	2.72	0.23	0.20	1.61	-	100.00
9	29.37	8.05	1.11	7.44	32.84	-	5.81	9.91	0.74	0.40	2.15	0.05	0.24	1.20	0.25	99.56
10	32.80	5.25	7.57	3.42	28.87	0.37	6.06	8.40	1.51	0.88	3.68	-	0.05	0.74	-	99.60
11	34.80	9.36	11.45	6.67	17.80	0.23	6.90	5.57	1.93	1.12	0.76	0.11	0.12	2.70	0.29	99.81
12	34.29	5.97	11.69	6.29	17.26	0.24	4.94	8.53	1.89	1.32	3.49	0.41	0.22	3.14	0.31	99.99
13	24.61	10.60	7.90	6.65	23.47	0.23	4.72	10.79	1.20	0.50	5.88	0.55	0.06	2.13	0.38	99.67
14	41.26	5.63	14.35	5.52	12.83	0.20	4.26	5.73	2.74	2.30	1.08	0.14	0.20	3.29	0.41	99.94
15	33.00	7.60	11.60	4.00	13.60	0.40	5.50	13.70	2.10	0.70	6.20	tr.	0.20	1.40	-	100.00
16	33.79	7.61	10.95	6.56	15.93	0.24	6.93	10.10	2.12	0.71	3.65	-	-	1.00	-	99.59
17	25.20	10.28	5.30	4.20	12.20	0.42	7.50	14.20	1.40	0.48	7.30	0.25	-	1.20	-	99.93
18	28.53	9.66	6.94	6.20	24.43	0.25	8.85	9.42	1.10	0.57	4.58	-	-	0.45	-	100.98
19	24.00	13.60	1.90	3.30	29.10	0.48	9.60	9.70	0.70	0.70	4.87	0.13	0.20	1.10	-	99.38

Chernigovka (Novo-Poltavka) massif: 1, 2 - ore alkaline pyroxenites; 3 – average composition. Oktyabrsky massif: 4, 5 - ore gabbro. Pokrovo-Kyreyevo massif: 6, 7 - ore pyroxenites. South-Kalchyk: 8, 9 - ore peridotites; 10 - ore gabbro. Davydky massif: 11 – troctolite; 12 - ore gabbro; 13 – ore cumulate; 14 - endocontact gabbro. Stremygorod deposit: 15, 16 - ore troctolites; 17, 18 - ore peridotites; 19 – ore cumulate.

Table 1. Chemical composition of some ilmenite- and apatite-bearing rocks from alkaline and subalkaline magmatic complexes of the Ukrainian Shield.

te, calcite and sphene. The ilmenite contents in these rocks vary between 5 and 15%, with an average at 10 % (Table 1). The ilmenite is enriched in Nb (up to 0.3 %). Some varieties of carbonatites from these complexes have high concentrations of ilmenite, which can thus become a useful accompanying mineral.

In the gabbro-syenitic complexes the ilmenite-rich rocks are represented by ore gabbros, troctolites, cumulate peridotites and pyroxenites. Two types of massifs are distinguished. In the first one, gabbro-syenitic massifs are associated with spatially and genetically related anorthosite-rapakivi granite plutons, e.g., the Davydky gabbro-syenitic massif, which lies in the northern part of the Korosten anorthosite-rapakivi-granite pluton (Fig. 1). The South-Kalchik massif (Azov area) represents a syenitic analogue of such plutons. In the second type of gabbro-syenitic complexes, ore gabbroids and pyroxenites are genetically related to compound polyphasial massifs (Oktyabrsky, Pokrovo-Kyreyevo), whose development terminated in nepheline syenites (mariupolites, foyaites, juvites).

Among the ilmenite-bearing gabbroids and peridotites of the gabbro-syenitic complexes, apatite-ilmenite-ore and essentially ilmenite-ore varieties are to be found (contents of ilmenite amount to 20%, and apatite to 5-10%). These rocks are rather similar to the ore troctolites of the Stremygorod ilmenite-apatite deposit, localized in the middle of the Korosten pluton. In our view, gabbroids of this deposit differ essentially from typical rocks of the Korosten pluton by their increased alkalinity (presence of titanaugite). All the above-mentioned deposits related to apatite-ilmenite gabbroids and peridotites have a magmatic origin and were formed mainly by crystal accumulation of ilmenite, magnetite and apatite. Their parental magmas were enriched in titanium and/or phosphorus, either at the mantle site of generation (nephelinites in the alkaline-ultrabasic complexes and alkaline basalts in the gabbro-syenitic types), or by a differentiation process in intermediate-level magma chambers.

The gabbro-anorthosite massifs of the Korosten Pluton (Ukraine) and problems concerning the evolution of the parental magmas

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Massive anorthosites represent one of the largest occurrences of Precambrian basic magmatism. Proterozoic anorthosite complexes contain numerous Fe-Ti deposits of economic grade and have therefore been studied extensively. The Korosten Pluton occurs in the northwest part of the Ukrainian Shield, in the Volyn Block. The pluton is a multiple anorogenic intrusion, composed of the classical association of gabbroid - anorthosite - rapakivi-granite, which intruded in Proterozoic times (1737-1800 Ma). The rocks of the Korosten Complex occupy an area of 12,000 km². The largest gabbro-anorthosite massifs are: Volodarsk-Volynsky (1250 km²), Chepovichsky (830 km²), Fedorovsky (104 km²), Pugachevsky (15 km²) and Krivotinsky (30 km²); they occupy 18% of the area. The massifs are flat, sub-horizontal bodies with thicknesses from 0.5 to 3 km. Gabbro-anorthosite varieties of basites predominate in most bodies. Anorthosites and mesocratic gabbroids are less abundant. Ultrabasites are rare. Anorthosites (gabbro-anorthosites) and gabbroids have long been considered as having been intruded in two phases (Polkanov 1948, Bukharev et al. 1973a,b). Recent detailed field studies have, however, shown that both anorthosites and gabbroids were intruding repeatedly (Mitrokhin 2000, Zinchenko et al. 2000]. Field, mineralogical, petrographical and geochemical investigations have brought new evidence showing that large gabbro-anorthosite massifs are multiple intrusions, composed of several age associations (suites) of basic rocks. In the Volodarsk-Volynsky, Chepovichsky, Fedorovsky and Pugachovsky massifs, five age suites of basic rocks are distinguished: Early Anorthosite Suite (A_1), Main Anorthosite Suite (A_2), Early Gabbroic Suite (G_3), Late Gabbroic Suite (G_4) and Dyke Suite (D_5).

The Early Anorthosite Suite, Main Anorthosite Suite and leucodolerites of the Dyke Suite form successive trends of sub-alkaline high-alumina basic rocks. Within each trend the decreasing plagioclase content with addition of mafic minerals is accompanied by the regular descent of SiO_2, Al_2O_3, CaO, Na_2O and increase of TiO_2, FeO and MgO. The ratios FeO:MgO and TiO_2:FeO within each suite are almost constant and correspond to fractionation and accumulation of plagioclase during ascent to the final level of emplacement. Within the high-alumina basite trend, transition from A_1 anorthosites - leuconorites to A_2 anorthosites -leucogabbros-norites and to D_5 leucodolerites can be traced in the variation of mineral associations and compositions: Pl_{42-57} + Opx_{22-46} to Pl_{45-62} + Pig_{44-54} + Aug_{38-45} ± Ol_{45-59} ± Fsp to Pl_{45-60} + Aug + Fsp ± Ol_{57-60}, with increasing contents of Fe-Ti oxides and apatite (the indices refer to 100*An/An+Ab in plagioclases and 100*Fe/Fe+Mg in mafics). This evolution is accompanied by decreasing contents of Mg, Cr, Ni and Sr and increasing contents of Ti, Fe, Mn, K, P, V, Sc, Ba, La, Ce, Zr, Y, Cu and Mo with a regular increase of FeO:MgO and TiO_2:FeO ratios.

The Early Gabbroic Suite, Late Gabbroic Suite and trachydiabase - trachyandesibasalts of the Dyke Suite form trends of sub-alkaline, moderately-aluminous, basic rocks. The major chemical evolution, within each suite, is marked by noticeable changes in FeO:MgO and TiO_2:FeO ratios. This is due to differentiation with separation of plagioclase and mafic minerals. From G_3 gabbro-norites to G_4 olivine gabbroids G_4 and to D_5 trachydiabases, the transition is Pl_{43-56} + Pig_{45-55} + Aug_{16-17} to Pl_{28-60} + Aug_{38-56} + Ol_{32-77} ± Pig_{56-64} ± Fsp and to Pl_{25-62} + Aug_{42-72} + Fsp, with abrupt addition of Fe-Ti oxides and apatite in G_4.

Taking into account the 'anorthosite scheme' of crystallisation for high-alumina basic suites it may be proposed that the

depletion in Mg, Cr, Ni and Sr and enrichment of Ti, Fe, Mn, K, P, V, Sc, Ba, La, Ce, Zr, Y, Cu and Mo with increasing $FeO:MgO$, and $TiO_2:FeO$, in row: (anorthosite -> leuconorite A_1) -> (anorthosite - leucogabbro-norite A_2) -> (leucodolerite D_2), are connected with the deep compositional evolution of the parental high-alumina basalt magmas. The Main Anorthosite Suite and Late Gabbroic Suite have inherited minerals and chemical compositions, which allows to consider them as cumulates (A_2) with a complementary residual liquid (G_4). The new data on the evolution of the Korosten basic magmatism point to the possible occurrence of Ti-Fe-P magmatic deposits in the G_4 and D_5 formations.

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First approach to the petrology of the Kamenka peridotite-gabbro-anorthosite intrusion

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The Kamenka layered intrusion was discovered in 1965 during geological surveying. Initially it was attributed to the Korosten AMCG complex, which occurs about 50 km to the southeast of Kamenka. However, strong differences in chemical composition between the leucocratic Kamenka basites and Korosten anorthosites were later recognised (Kogut et al. 1992), and some characteristics in the Kamenka whole-rock chemistry were recently used (Zinchenko et al. 1998) to argue for an affiliation with the huge Paleoproterozoic continental flood-basalt (CFB) province.

The Kamenka peridotite-gabbro-anorthosite layered intrusion is situated on the northern edge of the Ukrainian Shield. It is assumed from geophysical data that Kamenka is a typical batholith with a total area of about 500 km² and a thickness of up to 1300-1500 m. The main part of the intrusion is fault-bounded and downthrown by about 500 m and overlain by Phanerozoic sedimentary rocks; only a small 2-3 km-wide southern strip occurs near the surface beneath a 10m-thick Quaternary sedimentary cover.

The Kamenka massif intrudes into c. 1.975 Ga granites (and their comagmatic altered basites) and into metamorphosed volcanic and sedimentary rocks. The age of the Kamenka intrusion, as concluded from thermo-ionic emission analyses on zircon is 1.98 Ga. The rocks of the Kamenka intrusion are fresh, with only olivine-rich cumulates moderately serpentinized.

A Bottom zone and a Main Layered zone have been distinguished within the Kamenka intrusion.

The *Bottom zone* is composed of melanocratic and mesocratic mafic rocks (ortho-

clase-bearing gabbro and peridotite, monzonitic gabbro, monzonite, syenite, etc.), with heterogeneous, quite often inequigranular (ataxite) fabrics. Thin granite veins are noted in places within these rocks. The usual thickness of the Bottom zone is about 10 m. These characteristics suggest a high degree of contamination by acidic country rock.

The *Main Layered zone* is composed of peridotite, plagioclase-bearing peridotite, olivine-bearing melanocratic gabbro and melanocratic troctolite, meso- and leucocratic olivine-bearing gabbroid and anorthosite. These rocks sequentially grade into each others and appear to constitute a natural line of differentiation. At least three layered units (lower, middle and upper) are recognised within the Main Layered zone.

A high Mg-number, calculated as $\#Mg=Mg^{2+}/(Mg^{2+}+Fe^{2+}+Fe^{3+}\times 0.9)$, high Al_2O_3 abundances and high contents, of Ni and Sr are characteristic for the Kamenka intrusion. By contrast abundances of TiO_2, CaO, Cu, Zn, V, Sc, Rb, Zr, Nb, Yb, and Y are relatively low compared with typical CFB.

The Kamenka massif formed as a result of crystallization of high-Al tholeiitic melt which belonged to the Paleoproterozoic CFB province. The average composition of the Kamenka intrusion shows that the melt underwent extended crystallization prior to emplacement into the ultimate chamber. Although the Mg# of the average melt was rather high (0.62), at least 10% of high-Mg olivine (Fo_{91}) must be added to it in order to bring it into equilibrium with mantle residua. Meanwhile, the high Al_2O_3/CaO ratio (2.15 on average) provides evidence of extended crystallization of clinopyroxene or clinopyroxene plus plagioclase (in addition to olivine) *en route* to the ultimate chamber. It is noteworthy that the Kamenka line of descent follows the differentiation trend defined by the Paleoproterozoic CFB dykes. This trend appears to be due to precipitation of an assemblage of $An_{92}+Fo_{79}$ in the proportion 1.5:1.0, which suggests shallow (up to 10 km) crystallization (Green 1970). Thus, at least a two-stage evolution must be invoked for the Kamenka intrusion. The first stage involved crystallization of high-Mg olivine along with clinopyro-

xene at depth, while the second stage involved precipitation of plagioclase plus olivine in a shallow intermediate chamber, and terminated with at least three injections into the Kamenka ultimate magma chamber.

Although a genetic link between the Kamenka intrusion and other intrusions of the Paleoproterozoic CFB province seems to be clear, the *origin* of the melt with the observed chemical features is still unexplained. Extended (up to 30-40%) partial melting of a dry pyrolite source at a depth of 70-100 km is usually accepted (Green 1970) for a tholeiite-picrite liquid that is believed to be primary for continental tholeiites. Known partition coefficients and assumed mantle composition (Frey et al. 1978) predict that the abundances of most trace elements should be much higher than observed. Supposed precipitation of an olivine plus clinopyroxene assemblage and possible contamination with crustal material *en route* can only increase the abundances of the most incompatible minor elements. Three melt composition models, which include 1) simple equilibrium 5% partial melting of undepleted mantle; 2) simple equilibrium 30% partial melting of depleted model 1 mantle, and 3) model 2, complicated by precipitation of 10% of olivine and contamination with production of 3% partial melting of lower crust, were calculated. Compared with the Kamenka melt composition all models reveal similar features: 1) the abundances of Ba, Sr, Co, and Ni in the Kamenka melt are much higher than predicted in the models; 2) abundances of Rb, Zr, Y, Sc, V, Zn, Cu, and Cr are much lower than predicted. Among them, only the high content of Ba and Sr can be explained by contamination with crustal material, whereas low Cr contents may be attributed to extraction of clinopyroxene or spinel. It should be noted here that the Sr content of the Kamenka rocks is at least a factor of two higher than that typical for other CFB rocks, which requires a special explanation. Neither incompatible elements, such as Rb, Zr, Y, Sc, V and Zn, nor compatible (Ni) or partly-compatible (Co) elements fit the models. It is especially difficult to explain the high Ni content, which may appear due to the presence of sulphur in the melt, coupled with an extremely low Cu abundance.

It seems to the author that no reasonable explanation can be found at present to fit the observed abundances of minor and major (especially high contents of Al and low contents of Ti and P) elements. More extended geochemical investigations are required.

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Ti, Mg, and O isotopic compositions in ilmenites from Ukrainian ore deposits as indicators of their crystallization conditions

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A Cameca IMS-4f ion microprobe has been used for investigation of ilmenites from different rocks of the Ukrainian Shield (gabbroids, alkaline pyroxenites and carbonatites). Ilmenite crystallizes in a wide range of physico-chemical conditions, which usually results in isotope fractionation of elements such as O, C, S, etc. Previous investigations had shown that fractionation of Ti isotopes occurs also in meteorites and in some terrestrial minerals (Fahey et al. 1987, Zinner 1989). It is especially common in meteorites (-70 up to +273 ‰ for $\delta^{50}Ti$) (Fahey et al. 1985). Terrestrial minerals have not been investigated to the same extent as space material.

A notable fractionation of Ti isotopes has been established in ilmenites from alkaline pyroxenites and carbonatites of the Chernigovka complex, Azov area (Fig.1). The present authors tentatively explain this fractionation through variations in the physico-chemical conditions of formation of the ilmenite-bearing rocks (temperature, fugacity of oxygen, alkalinity). In the example of two pairs of ilmenites from carbonatites and alkaline pyroxenites, an increase in heavy (^{49}Ti and ^{50}Ti) isotopes was discovered in rocks formed in more alkaline and oxidizing conditions. It was also supposed that the fractionation occurred at a lower temperature than the primary crystallization temperature (accepted at about 1000°C). Hence, oxidation processes and increase of alkalinity of the magma, and, perhaps, decreasing temperature, promote enrichment of Ti heavy isotopes.

Ti isotopic compositions were also measured in individual ilmenites from gabbroids of the Korosten pluton. In these samples, $\delta^{49}Ti$ and $\delta^{50}Ti$ have rather insignificant values (0 and 1.5 ‰, respectively). They are similar to ilmenites from carbonatite (N 1). Microprobe analyses of ilmenites from the Korosten pluton and the Stremygorod apatite-ilmenite deposit show that they are also characterized by rather low contents of the hematite component (0.5-1.9 % Fe_2O_3). This testifies to rather reduced conditions of crystallization.
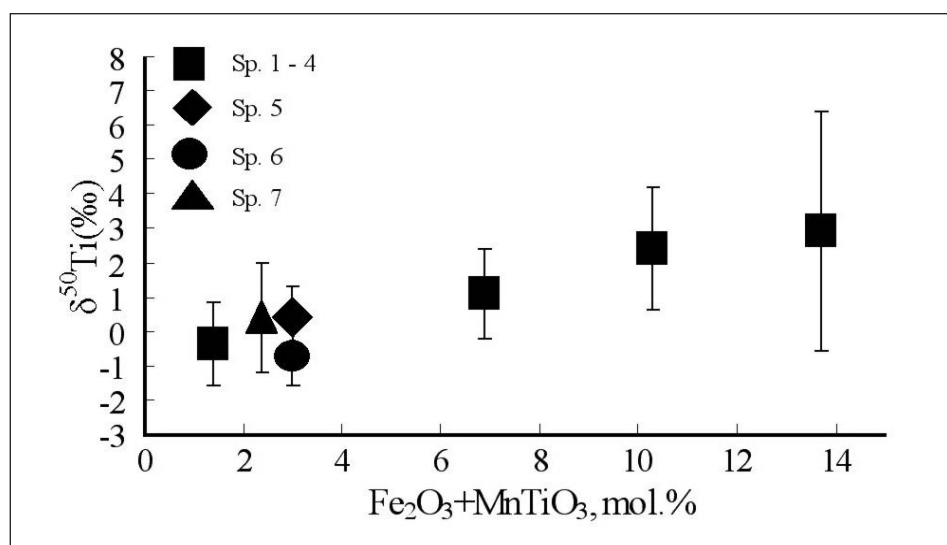
The tendency for an increase in the weight of O isotopes was formerly established in carbonates from carbonatites of the Chernigovka complex (Kryvdik et al. 1997). It has been shown that an increase in the Fe^{3+}/Fe^{2+} ratio in a rock is correlated to increasing ^{18}O values in carbonates, from 5 up to 17.5 ‰. $\delta^{18}O$ in magnetite from these carbonatites (+0.3 up to +5.0 ‰) indicates an equilibrium isotope distribution between carbonates and oxides (magnetite, ilmenite) at rather high PT values. The isotope distribution of carbon between carbonates and graphite in the same rocks also suggests a high temperature of equilibrium crystallization of these minerals.

O isotopic compositions (conventional mass spectrometry) of ilmenites from alkaline pyroxenites show a narrow range of $\delta^{18}O$ from +1.96 up to +2.98 ‰, with a trend of increasing $\delta^{18}O$ with increasing alkalinity and oxidation of the rocks, as already mentioned above for Ti isotopes in ilmenites from the same rocks ($\delta^{50}Ti$).

In conclusion, Ti, Mg and O isotopic compositions in ilmenites depend on the rock type and on the crystallization conditions (temperature, alkalinity, oxygen fugacity). Our investigations point to oxygen fugacity as being one of the primary factors controlling the isotopic composition of magmatic ilmenite.

Fig.1. Dependence of $\delta^{50}Ti$ versus (Fe_2O_3 + $MnTiO_3$) contents in ilmenites from carbonatites and alkaline pyroxenites (Sp. 1-4) of the Chernigovka complex, gabbroids of the Korosten pluton (Sp. 5), ore troctolites of the apatite-ilmenite Stremygorod deposit (Sp. 6) and Lunar rocks (Sp. 7) (Fahey et al. 1987).

Mg isotopic compositions were determined in ilmenites from basic and ultrabasic rocks of the Korosten pluton and related complexes (Gorodnytza, Prutovka and Varvarovka). The ultrabasic rocks of the Gorodnytza intrusion are alkaline (jacupirangites and melteigites) and considered to be of deep (mantle) origin ($\delta^{26}Mg=2\pm1$ ‰). The Prutovka complex belongs to a differentiated intrusion, the basic and ultrabasic rocks (early differentiates) of which are nickel-bearing. This complex belongs to a Precambrian trap formation. The Varvarovka complex is supposed to have been significantly contaminated by crustal granitoid material. Starting from these characteristics, we explain the various Mg isotopic compositions in ilmenites from these complexes. Thus, the $\delta^{26}Mg$ in ilmenite from the Prutovka complex is close to that in diabases, and the $\delta^{26}Mg$ in ilmenites from the Varvarovka complex is nearer to those in granites. The data obtained in the Prutovka complex suggest a definite dependence of $\delta^{26}Mg$ values on the degree of oxidation of iron in ilmenites (Fe_2O^3 contents).

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Ti, V, Pt, Pd and Au in Travyanaya Bay ore peridotites and their possible genetic relation with Belomorian Mobile Belt anorthosites

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Ore peridotites are exposed near Travyanaya Bay on Lake Keret, North Karelia (Russia), in the Belomorian Mobile Belt (BMB), eastern Fennoscandian Shield, as a part of the differentiated Palojarvi massif. Mineralization, composed of ilmenite and magnetite with, locally, traces of pyrite and chalcosine, occurs in host ore rocks such as harzburgite, lherzolite, websterite and amphibolite. Harzburgite shows a sideronite texture and consists of 25 % olivine (Fa_{41}), 21% orthopyroxene (Fs_{34}), 41% magnetite and 10% ilmenite and with secondary minerals such as hornblende (f = $Fe^{2+}+Fe^{3+}/Fe^{2+}+Fe^{3+}+Mg=$ 0.25), chlorite, serpentine, garnet, carbonate and spinel. Augite is commonly present in lherzolite and websterite. These rocks show gradual transitions and, obviously, form a continuous differentiated sequence. During amphibolization they were converted into ore amphibolites, most similar in chemical composition to websterite.

The valuable components of the ore rocks are represented by TiO_2 (3.0-7.4 wt.%), V_2O_5 (0.21-0.36 wt.%), Pt (up to 0.52 ppm), Pd (up to 2.8 ppm) and Au (up to 2.5 ppm). Ilmenite concentrates Ti, and magnetite, clinopyroxene and amphibole concentrate V (in decreasing order). Separate mineral phases of Pt and Pd are unknown, but the maximum contents of these elements are observed in olivine and ilmenite. Au contents are maximum in ilmenite and much smaller in magnetite and oli-vine. The average amount of Pt + Pd is 1.23 ppm in ore peridotite and olivine websterite, 1.48 ppm in amphibolized websterite and 1.022 ppm in ore amphibolite.
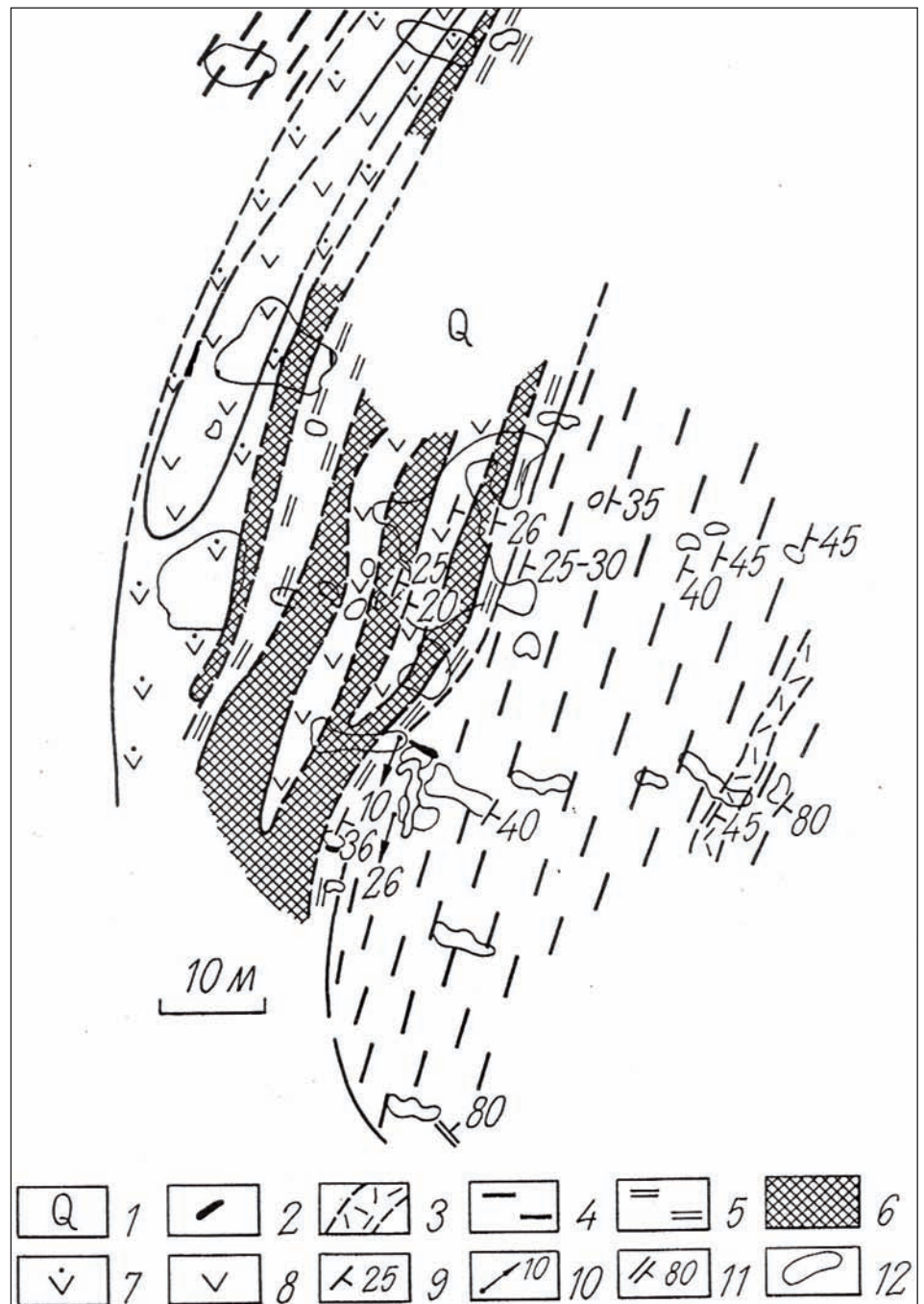

Fig 1. Geological map of the "Guba Travyanaya" area. Legend: 1 – quaternary rocks; 2 – pegmatitic veins ; 3 – leucocratic, mostly plagioclasic amphibolites; 4 – garnet- and garnet feldspathic amphibolites; 5 – ore amphibolites (Opx & Cpx); 6 – ore websterites; 7 – ore peridotites with microareas of symplectites; 8 – ore peridotites; 9-11 – structural elements ; 12 – outcrops

Ore rocks form small bodies in a Fe-rich (f > 0.5) garnet-feldspathic amphibolite zone. About 10 bodies composed of ore rocks have been discovered. The largest and most thoroughly studied body has a visible thickness of 30 m, dips steeply east and has been traced for 100 m along strike (Fig. 1). The boundary between the body and plagio-amphibolite extends across a 30-40 cm-thick thin-laminated zone formed by compositionally variable amphibolite. The ore rock-body wedges out northwards and southwards. The other ore rock- bodies are much smaller.

The association between ore rocks and garnet plagio-amphibolites is interpreted as a metamorphosed differentiated series in the fairly large (about 8 km²) Palojärvi massif, which has been tectonically divided into three large fragments, markedly differing in chemical composition. The largest, western fragment is built up by garnet-feldspathic amphibolite, with zones exhibiting well-preserved gabbro structures. The eastern fragment is structurally similar, but its relict zones contain both gabbronorite and plagioperidotite. The southern fragment is richer in Fe (f > 0.5) than the western and eastern fragments, and is presumably composed of late differentiates. Ore peridotites have only been found in this fragment, and are interpreted as melanocratic cumulates.

Leucocratic plagio-amphibolite, similar to apoanorthositic rocks, occur in the eastern portion of the fragment.

The Palojärvi massif thus comprises differentiated rocks from Mg-rich plagioperidotite (eastern fragment) to rocks similar to ferrogabbro and ore peridotite. An intermediate position is occupied by mesocratic and leucocratic amphibolites (f < 0.5), corresponding in chemical composition to gabbro and leucogabbro.

Comparison of the Palojärvi rocks with other BMB intrusive rocks shows that the ore rocks have no counterparts in the province. The Mg-rich portion of the massif is most similar to the mesocratic units in the gabbro-anorthosite complexes (Nizhnepopovsky, Boyarsky, Severopezhostrovsky and other massifs). Both gabbro-anorthosites and Palojärvi rocks are intensely metamorphosed and foliated; therefore, gabbroids pass into plagio-amphibolites throughout the studied area. Ilmenite, which is characteristic of Travyanaya Bay rocks, is a major ore mineral in gabbro-anorthosites.

To sum up, ore rocks similar to those in Travyanaya Bay are likely to be found in connection with gabbro-anorthosite massifs elsewhere in the BMB.

Precambrian anorthosites in the Belomorian Mobile Belt, eastern Fennoscandian Shield

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In the Belomorian Mobile Belt (BMB) (western White Sea region), anorthosites are known in at least three geological settings:

1. The most strongly differentiated leucocratic portion of an Early Proterozoic (2.4 Ga), drusitic, lherzolite-gabbronorite complex (Stepanov 1981) is represented by anorthosites. They are subordinate and form small lenses up to tens of metres in thickness and schlieren in olivine gabbronorites and plagioclase lherzolites. Massifs of this complex have primary intrusive contacts with chilled zones (1.5-2 m), apophyses and enclosed rock xenoliths. In addition to these intrusive forms, dykes are also characteristic of the complex. The parent magma of these intrusions corresponds to high-MgO, high-Cr tholeiite, rich in SiO_2, determined as a mean of chilled margins with the following composition (wt.%, ppm, n=31): (50.13 SiO_2, 0.69 TiO_2, 11.41 Al_2O_3, 1.59 Fe_2O_3, 9.02 FeO, 0.19

MnO, 14.89 MgO, 8.61 CaO, 1.70 Na_2O, 0.53 K_2O, 0.07 P_2O_5, 0.11 H_2O, 172 Cr, 80 Ni, 47 Co, 243 V, 4 Zr, 9 Y. This type of anorthosite is described as stratiform.

2. Anorthosites are an essential component of gabbro-anorthosite massifs (gabbro-anorthosite complex; Stepanov 1981). In these bodies, anorthosites as well as leucocratic and mesocratic gabbro strongly dominate over melanocratic rocks that display a cumulate structure and commonly occur in the lower part of the intrusion. Judging by the composition of chilled rock autoliths in the Boyarsky complex, the parent magma of this complex seemed to correspond in composition to high-alumina gabbro with (wt.%, ppm) 52.35 SiO_2, 0.17 TiO_2, 16.30 Al_2O_3, 1.82 Fe_2O_3, 5.62 FeO, 0.16 MnO, 9 MgO, 12.02 CaO, 2.28 Na_2O, 0.15 K_2O, 0.03 P_2O_5, 0.11 H_2O, 172 Cr, 80 Ni, 47 Co, 243 V, 4 Zr and 9 Y .

The isotopic age of the gabbro-anorthosites is 2450 Ma in the Kolvitsa massif (Mitrofanov et al. 1993) and 2452 Ma in the Severopezhostrovsky massif (Alexejev et al. 2000). These ages show that their formation was as close as possible in time to the formation of the lherzolite-gabbronorite complex. Based on this evidence, some authors combine these units in one single drusitic complex. However, the above groups of intrusion (complexes) differ substantially in the composition of parent melts and in relative geological ages. Dykes of the lherzolite-gabbronorite complex cut the gabbronorite massifs. The latter are typically strongly reworked. Their primary contacts are unknown. Intrusive rocks usually pass gradually into banded garnet amphibolites. Lherzolite-gabbronorite bodies have crosscutting contacts with similar amphibolites. Therefore, differentiated gabbro-anorthosite massifs should be more closely studied isotopically.

3. Anorthosites also occur in thin (several metres to tens of metres) lenticular beds in banded amphibolites that are widespread in the BMB and form thick horizons. Anorthosites are typically coarse grained and foliated. Their boundaries with hosting amphibolites have no magmatic relationships, suggesting that anorthosites are the easiest recognizable fragment of a differentiated series in which the gabbroid portion is fully transformed into amphibolites. Small anorthosite bodies, rimmed

by alumina-rich feldspathic amphibolites, occur near Lake Skalnye, Lake Nigrozero, Kivguba and Point Kartesh in the White Sea and in other parts of the BMB. The largest massif of this type is the crescent-shaped Kotozero massif, with a visible thickness of 1.5-2 km. Its axial part is composed of foliated meta-anorthosites and less common massive anorthosites with relics of magmatic structure. The most strongly altered anorthosite in the Kotozero massif has the following composition (wt.%, ppm): 50.90 SiO_2, 0.25 TiO_2, 29.04 Al_2O_3, 0.75 Fe_2O_3, 1.32 FeO, 0.03 MnO, 0.61 MgO, 12.76 CaO, 3.53 Na_2O, 0.21 K_2O, 0.06 H_2O, 0.42 LOI, 0.05 P_2O_5, 15 Cr, 28 V, 32 Ni, 16 Co and 40 Cu. On the periphery, the foliated anorthosites are rimmed by a feldspathic amphibolite band – presumably a mesocratic constituent of the massif. Amphibolites consist of 49.4 SiO_2, 1.20 TiO_2, 15.77 Al_2O_3, 3.42 Fe_2O_3, 8.62 FeO, 0.20 MnO, 6.87 MgO, 10.15 CaO, 2.15 Na_2O, 0.14 K_2O, 0.07 H_2O, 0.16 P_2O_5, 171 Cr, 111 Ni, 63 Co, 280 V. Occurring in meta-anorthosites are dykes of fine-grained norites with the composition (wt.%, ppm) 50.86 SiO_2, 0.56 TiO_2, 10.93 Al_2O_3, 1.37 Fe_2O_3, 9.08 FeO, 0.17 MnO, 15.22 MgO, 0.83 CaO, 0.15 Na_2O, 0.42 K_2O, H_2O < 0.07, 0.91 LOI, 0.09 P_2O_5, 1183 Cr, 190 V, 150 Ni, 63 Co and 80 Cu and garnet-diopside-plagioclase rocks with the composition 51.40 SiO_2, 1.90 TiO_2, 12.75 Al_2O_3, 2.88 Fe_2O_3, 12.93 FeO, 0.22 MnO, 4.35 MgO, 8.70 CaO, 1.46 Na_2O, 1.04 K_2O, 0.10 H_2O, 1.54 LOI, 0.28 P_2O_5, 102 Cr, 302 V, 79 Ni and 120 Cu, and possibly dykes of a complex composed of lherzolites, gabbronorites and garnet gabbro. They are deformed together with anorthosites, but the degree of deformation is lower in the dykes than in the rocks of the massif.

The structural line traced by the Kotozero massif and by the smaller gabbro-anorthosite bodies that lie on its extension, seems to correspond to a dislocation zone which evolved in an extensional regime in Late Archaean and Early Proterozoic times, and in a compressional regime in Svecofennian times. The close relationship between these gabbro-anorthosites and amphibolites appears to reflect their Archaean age. At the same time, the extent of amphibolitization is not a major criterion to distinguish them from the gabbro-anorthosite massifs described above (point

2). Based on this evidence (Stepanov 1981, Stepanov & Slabunov 1989), they were formerly described as part of one complex. The isotopic age of this variety of gabbro-anorthosites has not yet been estimated. They are morphologically similar to autonomous anorthosites (massive type).

4. The formation of the lherzolite-gabbro-norite complex is commonly attributed to Early Proterozoic rifting. The position of gabbro-anorthosite magmatism in the BMB is more obscure. We (Stepanov & Slabunov 1989) assumed its formation as a manifestation of Late Archaean reactivation (post-collisional stage of evolution). The available isotopic dates, indicating an Early Proterozoic age for several massifs typical of the complex, have led us to conclude that some of the generally accepted views should be revised and that the complex should be investigated more thoroughly.

5. Associated with the gabbro-anorthosite massifs of the BMB are the occurrences of Au-Pt mineralization in two parageneses: 1) in association with sulphide mineralization (Ileiki Islands, White Sea) and 2) in association with ilmenite-magnetite mineralization (Travyanaya Guba ore peridotites).

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Subduction-related leucogabbro-anorthosite ilmenite-bearing series: an example of water-rich high-temperature anatexis, platinum belt of the Urals, Russia

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The platinum belt of the Urals is composed of zonal massifs belonging to the Ural-Alaskan type of the dunite-clinopyroxenite-gabbro rock series (420-430 Ma), with famous platinum and magnetite deposits related to dunites and clinopyroxenites - hornblendites, respectively. The belt is assumed to be associated with a Late Ordovician-Silurian subduction zone dipping to the east. In the eastern part of the belt a suite of hornblende leucogabbro, anorthosite and plagiogranite dykes occurs in some zonal massifs together with a leucogabbro-anorthosite-plagiogranite, the Chernoistochinsk massif, which is described below.

The root zone is composed of migmatized hornblende gabbro and is exposed in the northern part of this massif. Migmatization (partial melting) has produced non-uniform rocks consisting of leucocratic patches full of gabbro relics and

Table 1: Major (wt.%) and trace (ppm) element composition of rocks from the Chernoistochinsk massif.

N	1	2	3	4	5	6	7	8	9	10	11	12	13
Sample	571	578	566	565	562	117	124	115	567	568	566	562	117
SiO_2	46,76	43,98	46,35	52,17	53,05	65,94	70,54	73,67	44,50	44,80	44,08	44,08	44,36
TiO_2	0,79	2,23	1,09	0,70	0,66	0,20	0,14	0,13	1,25	1,28	1,28	1,55	0,99
Al_2O_3	12,98	15,85	17,14	22,90	23,16	18,59	15,98	14,19	8,27	7,92	10,48	10,46	8,70
Fe_2O_3	6,18	7,41	4,87	4,00	2,91	0,99	0,90	0,56	8,65	8,36	5,61	5,67	5,99
FeO	6,46	8,61	7,67	2,15	3,88	1,53	1,76	2,04	10,41	11,13	11,61	10,75	12,57
MnO	0,24	0,26	0,21	0,10	0,09	0,03	0,03	0,03	0,39	0,39	0,30	0,48	0,42
MgO	8,03	4,99	5,11	2,10	1,85	0,70	0,52	0,28	10,75	10,80	11,08	11,31	11,73
CaO	11,96	10,06	11,72	8,37	8,26	6,08	5,44	4,28	11,09	11,19	11,38	11,05	11,12
Na_2O	3,08	2,69	3,10	6,16	4,97	4,78	3,73	3,78	1,39	1,39	1,52	1,24	1,22
K_2O	0,12	0,12	0,12	0,10	0,14	0,08	0,12	0,08	0,16	0,18	0,20	0,20	0,25
P_2O_5	0,07	0,28	0,22	0,28	0,26	0,08	0,03	0,04	0,24	0,09	0,02	0,06	nd.
LOI	2,47	1,23	1,61	0,63	0,76	2,04	1,94	1,68	1,77	1,83	2,39	2,46	nd.
Li	2,44	2,20	1,83	2,07	2,37	0,00	0,00	0,00	2,18	1,31	13,52	0,00	1,74
Rb	0,63	0,35	0,00	0,00	0,00	0,13	0,11	0,04	0,38	2,00	1,91	0,09	0,31
Be	0,63	0,69	1,00	0,93	0,98	1,13	0,82	0,92	0,71	0,72	0,80	0,66	0,66
Sr	493	611	1366	1250	1515	725	768	763	166	94	128	139	93
Ba	64	71	110	119	126	88	114	103	48	66	128	65	56
Sc	67	62	9	10,3	9,6	8,6	5,4	4,6	50	57	58	65	64
V	362	475	145	179	155	66	52	26	326	406	468	451	500
Cr	72	1	3	2,2	9,2	2,6	1,7	82,9	729	704	34	69	22
Co	46	46	8	9,9	8,4	4,8	4,6	3,4	56	54	43	45	58
Ni	29	5	2	5,9	3,8	51,4	65,6	77,6	260	241	18	21	42
Cu	78,2	159,1	54,0	57	49	58	21	26	190	69	52	23	96
Zn	112,4	176,5	45,5	68,3	56,7	6,0	9,3	3,8	195	198	253	259	219
Ga	16,1	22,3	23,2	24,0	24,5	14,8	12,4	10,6	19	20	26	25	21
Y	21,8	25,2	6,3	9,0	7,7	1,5	1,3	0,8	42,6	47,7	38,3	48,1	36,2
Nb	1,51	4,18	0,95	1,20	1,36	0,21	0,23	0,21	5,40	3,26	4,23	2,59	4,33
Ta	0,11	0,28	0,07	0,08	0,09	0,28	0,36	0,26	0,69	0,19	0,05	0,00	1,71
Zr	20,06	17,12	6,11	6,21	7,09	1,56	0,69	0,30	42,5	38,1	57,5	40,6	27,0
Hf	0,70	0,71	0,27	0,27	0,23	0,14	0,07	0,04	1,56	1,58	2,40	1,68	1,26
Pb	3,18	7,90	3,46	4,29	4,01	0,00	0,23	0,00	4,14	4,14	10,04	4,26	10,55
U	0,02	0,03	0,00	0,00	0,04	0,00	0,00	0,00	0,06	0,10	0,49	0,00	0,03
Th	0,05	0,06	0,03	0,03	0,07	0,00	0,00	0,00	0,00	0,96	4,21	1,42	0,09
La	4,33	5,98	2,74	3,76	3,21	0,72	1,24	0,62	6,70	6,79	11,18	8,00	3,97
Ce	12,69	16,66	6,61	9,80	8,95	1,71	2,25	1,15	27,47	26,78	33,57	32,22	16,38
Pr	2,14	2,76	1,04	1,59	1,49	0,23	0,23	0,15	5,42	5,31	5,73	6,85	3,56
Nd	10,31	13,29	5,01	7,53	7,05	1,16	1,13	0,67	27,08	26,54	30,22	40,49	19,41
Sm	2,81	3,73	1,23	1,90	1,78	0,33	0,32	0,14	7,26	7,39	8,70	11,38	5,90
Eu	1,10	1,59	0,73	0,94	0,89	0,23	0,24	0,14	2,56	2,43	2,53	2,91	1,84
Gd	3,08	4,07	1,30	1,90	1,72	0,36	0,25	0,15	7,29	7,61	6,96	7,69	6,03
Tb	0,52	0,65	0,19	0,27	0,25	0,05	0,04	0,02	1,13	1,23	1,03	1,35	0,96
Dy	3,60	4,40	1,11	1,62	1,31	0,30	0,25	0,15	7,51	8,08	6,89	8,66	6,45
Ho	0,79	0,98	0,23	0,34	0,29	0,07	0,05	0,03	1,57	1,77	1,46	1,79	1,36
Er	2,14	2,62	0,62	0,88	0,79	0,16	0,10	0,08	4,34	4,78	3,76	4,81	3,63
Tm	0,33	0,39	0,09	0,13	0,11	0,02	0,01	0,01	0,64	0,72	0,54	0,69	0,53
Yb	2,19	2,49	0,53	0,79	0,67	0,15	0,13	0,08	4,06	4,81	3,45	4,21	3,44
Lu	0,31	0,35	0,08	0,11	0,09	0,02	0,02	0,01	0,56	0,66	0,49	0,60	0,49

1, 2 - hornblende gabbro; 3 - partially melted gabbro; 4 - leucogabbro; 5 - anorthosite, 6 - quartz anorthosite; 7, 8 - plagiogranites; 9, 10 - hornblendites; 11-13 - hornblende from gabbro (11), anorthosite (12) and plagiogranite (13).
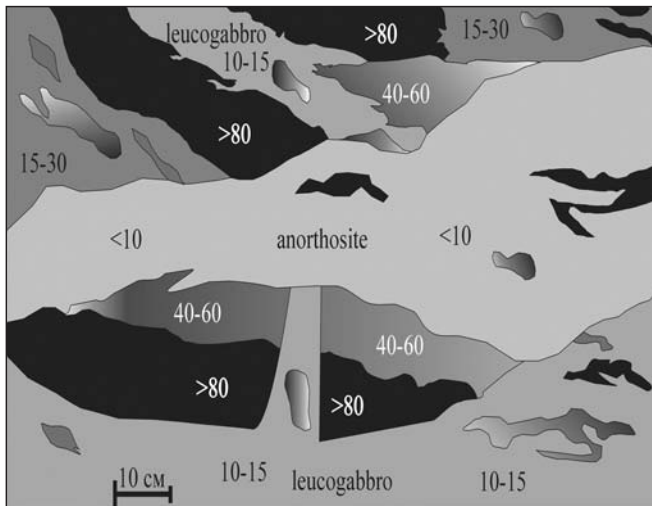
Fig. 1. Detail from a outcrop of migmatite; migmatized gabbro with anorthosite vein (light) and hornblendite restite (black). The numbers indicate the percentage of hornblende in the rock.
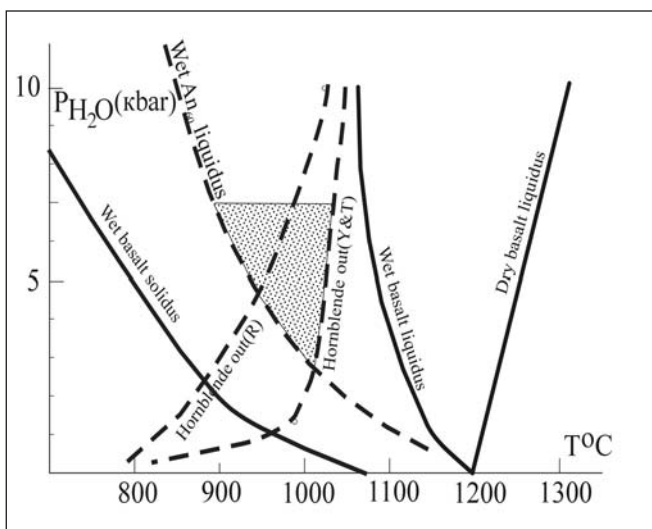

Fig. 2. P-T conditions of anatexis. Wet solidus and wet liquidus as well as An_{60} liquidus are represented for high-Al basalt (Yoder and Tilley, 1962), which is close in composition to the hornblende gabbro of the Chernoistochinsk massif. Amphibole stability lines are after Yoder & Tilley (1962) and Rushmer (1991). Dashed area represents the assumed field of anatexis.

hornblende segregations. Direct field investigations show that leucocratic segregations forming the hornblende leucogabbro represent the initial anatectic melt. Crystallization of this melt produced relatively homogeneous veins and small intrusive bodies in partially melted hornblende gabbros. Crystal fractionation of this leucogabbro melt gave rise to a leucogabbro-anorthosite-plagiogranite rock series mainly forming veins with gabbro xenoliths (Fig. 1). The hornblendite inclusions and schlieren represent the restitic phases. All the rocks have the same mineralogy, which is as follows: hornblende + plagioclase + magnetite + ilmenite. Magnetite is the main opaque mineral in gabbro, while ilmenite is predominant in anorthosite. Quartz appears in plagiogranite. The volume of these rocks is very small, not more than 3-5%. The amount of hornblende and the An contents in plagioclase decrease from gabbro to granites from 50 to 3-5 % and from 45-50 to 25-30%, respectively. The hornblende contents are the best visible index of partial melting and the degree of crystal fractionation, as shown in Fig. 1. The range of chemical compositions of the rock series (Table 1) represents the products of partial melting reactions from protolith (anal. 1 - 3) to initial melt (anal. 4) and products of its fractional crystallization (anal. 5-8) to restitic phases (anal. 9,10). Partial melting produces rocks impoverished in all the elements which are concentrated in femic minerals, such as Mg, Fe, Ti, and in most rare elements, with the exception of Sr, Ba and some others related to plagioclase. The magnitude of the positive Eu anomaly clearly increases in the range from gabbro to plagiogranite. At the same time, the ilmenite/magnetite ratio increases from 0.5 in gabbro to 10 in anorthosite.

The conditions of anatexis are shown in Fig. 2. This took place in the field of hornblende stability, while plagioclase with an anorthite content of less than An60 has undergone melting (shaded area in Fig. 2). Mass-balance calculations show that major- and trace-element compositions of leucogabbro correspond to approximately 60% of partial melting of hornblende gabbro. It is necessary to melt about 10% hornblende and nearly 100% plagioclase to produce the leucogabbro composition. A large body of experimental and geological work confirms the well-known fact that the usual quartz-bearing composition of anatectic melts has resulted from partial melting of amphibolite. The mechanism discussed above can occur when high temperatures (maybe because of a short time interval between gabbro intrusion and anatectic events) are combined with high concentrations of water coming from the subducted slab. It is believed that such quartz-absent anatexis was common in the early stages of the Earth's evolution and that it can be the source of some autonomous anorthosites.

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The Riphean layered intrusions of the Western Urals and related ilmenite-titanomagnetite deposits

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The Riphean layered gabbro intrusions are localized inside the graben facies of sedimentary and volcanic rocks of the paleo-continental sector of the South Urals. Volcanic rocks overlie the gabbro intrusions and are represented mainly by tholeiitic basalt and by dykes and small intrusions of rhyolite. The intrusions are underlain by sedimentary carbonate rocks and contain xenoliths of both sediments and volcanics. The Rb-Sr age of the volcanic rocks corresponds to 1346±41 Ma, and the initial ratio $^{87}Sr/^{86}Sr$ is 0.7098±0.001 (Krasnobaev & Semikhatov 1986). The gabbros have yielded an age of 1300 ±42 Ma and the initial ratio $^{87}Sr/^{86}Sr$ is 0.7093±0.0067 (Gorozhanin 1998).

Gabbro intrusions form a 70 km-long belt consisting of 4 isolated bodies dipping at 60° to the east. In the two southern massifs (Matkal and Kopan), gabbros are represented by the two-pyroxene variety (gabbro-norites) and contain high-Ti titanomagnetite ore bodies up to 10-15 km long and 1-2 m thick, parallel to the gabbro layering. These massive ores are surrounded by an aureole of impregnation. The northern massifs (Medvedevka and Kusa)

consist of hornblende gabbros and gabbro-amphibolites in the upper part and ilmenite gabbro-norites in the lower part (Table 1).

The ore bodies are concentrated in the gabbro-amphibolites and consist of ilmenite or low-Ti titanomagnetite+ilmenite. Ilmenite ores mainly belong to the impregnation type, while Ti-magnetite+ilmenite ores form both impregnation and massive types (Table 2).

The first ore type is situated mainly in the upper part of the massifs, the second type in the central part. Most Ural geologists explain the clear differences in composition of gabbros and related ores between the southern and northern group of massifs by the metamorphic overprint which has affected the northern group.
Our data, however, suggest that these differences are due to different P-T conditions of formation. Rocks and ores of the southern group were formed at a depth of 4-6 km (P_{tot} = 1-2 kb, P_{H2O}=0.5P_{tot}, T°C=900°-1100°, fluid enriched in F), while in the northern group these parameters are as follows: depth of 20-25 km,

Massif	SiO$_2$	TiO$_2$	Al$_2$O$_3$	Fe$_2$O$_3$	FeO	MnO	MgO	CaO	Na$_2$O	K$_2$O	P$_2$O$_5$	LOI	Total
Matkal (3)	46.91	2.19	14.89	3.89	8.57	0.11	6.66	11.21	3.00	0.42	0.06	1.58	99.49
Kopan (29)	44.50	2.90	15.47	5.72	10.25	0.09	5.66	11.4	2.36	0.34	0.14	0.57	99.64
Medvedevka (11)	44.81	2.26	14.69	5.25	11.76	0.17	7.76	9.34	2.40	0.50	0.07	0.89	99.90
Kusa (21)	43.90	3.14	13.24	5.61	11.17	0.19	7.07	11.54	1.85	0.22	0.23	1.41	99.57
Kusa, ilmenite gabbro-norite (11)	42.20	6.07	13.11	5.01	13.29	0.24	6.13	9.49	2.14	0.33	0.16	1.87	100.04
Kusa, gabbro-amphibolite (19)	46.44	2.16	16.26	4.57	9.16	0.15	5.53	10.66	2.96	0.43	0.06	1.74	100.12

Table 1. Average chemical compositions of the gabbros

Element	Ilmenite type		Titanomagnetite type	
	South group	North group	South group	North group
Fe	14-15	14-15	22-23	23-24
TiO$_2$	6-7	7	7	6-7
V$_2$O$_5$	0.12	0.12	0.25	0.25

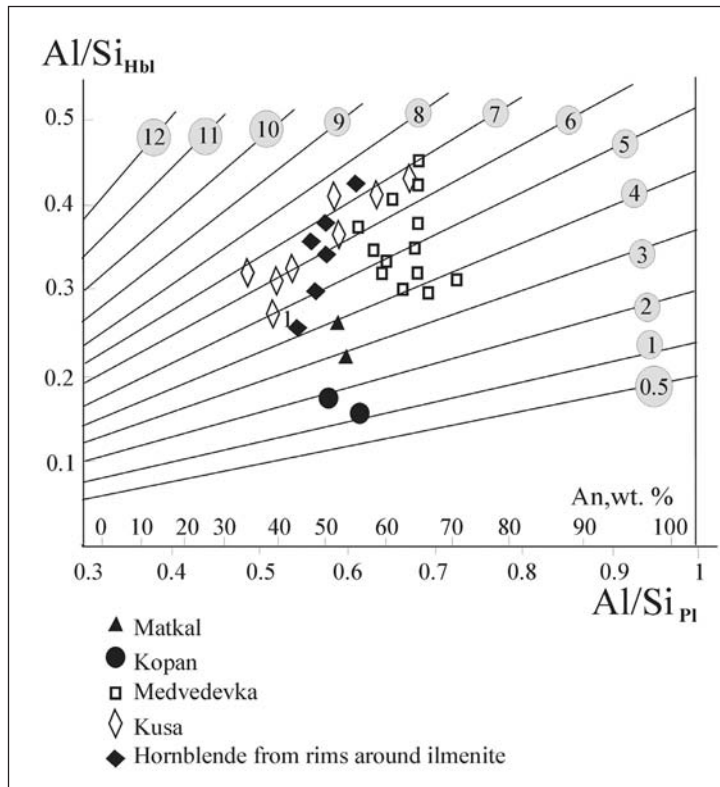Table 2. Average chemical compositions of the ores (wt.%)

Fig. 1. Plot of Al/Si in hornblende versus Al/Si in plagioclase for determination of P_{H2O} conditions of plagioclase-hornblende equilibrium. Numbers in circles are the pressure in kb (after Fershtater 1990).
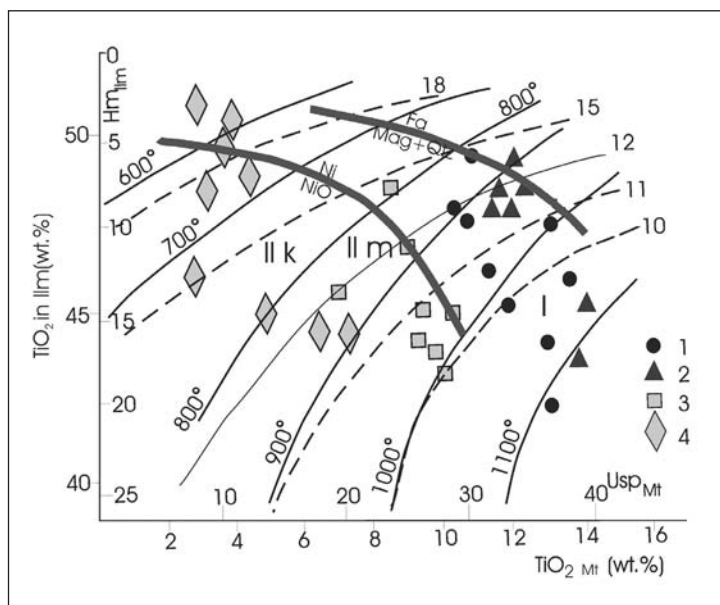
Fig. 2. Recalculated Buddington-Lindsley thermometer. 1- 4 symbols mark the magnetite-ilmenite pairs from the same massifs as in Fig. 1. Solid lines represent the temperature (°C) and dashed lines mark log of fO_2 (bars).

P_{tot} = 8-4 kb, P_{H2O}=0.7-0.9P_{tot}, T°C=850°-650°, fluid enriched in Cl (Figs. 1,2). Fluid enrichment in the massifs of the northern group is responsible for hornblende gabbro crystallization and its postmagmatic transformation into gabbro-amphibolites. Low temperature has caused the separate crystallization of ilmenite and low-Ti titanomagnetite, that has led to the formation of the different are deposits.

The upper part of the Kusa deposit was mined before the 1990s. Proven reserves of both ore types in the Matkal, Kopan and Medvedevka deposits range up to 7 billion tons.

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Rogaland Anorthosite Province – Syntheses

Introduction to the geological map of the Rogaland Anorthosite Province 1:75,000

Mogens Marker, Henrik Schiellerup, Gurli Meyer, Brian Robins & Oliver Bolle

Preamble

The geological map (Plate 1) has been compiled from available sources (see list on geological map) supplemented with mapping and field checks made by the Geological Survey of Norway during the summers 1997-2002. Key contributions for the map are provided by Michot (1960), Michot (1961), Michot & Michot (1969) and Rietmeijer (1979). For the Åna Sira anorthosite massif the mapping by Krause et al. (1985) has been used. Their work was compiled and digitised by Karlsen et al. (1998). Robins and Wilson (unpublished material) provided data for the Bjerkreim-Sokndal Layered Intrusion, Bolle (1998, unpublished thesis) for the Apophysis of the Bjerkreim-Sokndal intrusion, Maquil (1980, unpublished map reproduced in Duchesne & Maquil (1987)) for the Egersund-Ogna anorthosite massif, and Michot & Michot (1969) and Demaiffe et al. (1973) for the Hidra anorthosite massif. Large parts of the Håland-Helleren anorthosite complex, originally investigated by Michot (1961), have been re-mapped (M. Marker, R.B. Edland) as have the outer zone of the Egersund-Ogna anorthosite massif and the surrounding gneisses up to the contact (M. Marker). The area of jotunitic to mangeritic intrusive rocks in the western-most part of the province northwest of Bjårvatnet has been re-investigated (H. Schiellerup), and the areas immediately south of the Bjerkreim-Sokndal Layered Intrusion have also been re-mapped (G. Meyer). The host-rock Sveconorwegian gneisses have been covered by new mapping between the Apophysis and Lunde-vatnet (south of Heskestad) in the eastern part of the area (M. Marker). The map data for the Sveconorwegian gneisses north of the Bjerkreim-Sokndal Layered Intrusion have been adopted from Hermans et al. (1975). All the re-mapping and numerous field checks throughout the province also aimed to ensure a precise transformation of data from an old to a modern topographic basis.

Subdivision of the Rogaland Anorthosite Province

The Rogaland Anorthosite Province comprises a number of massif-type anorthosite bodies, the Bjerkreim-Sokndal Layered Intrusion and a range of other noritic and jotunitic to charnockitic intrusions. All units were emplaced into Sveconorwegian gneisses that were deformed under granulite-facies conditions. The subdivision into rock bodies (see inset map on the geological map) generally follows the previously established subdivision of the province. An exception is the Håland-Helleren massif, which has traditionally been treated as a single anorthosite unit. This massif has now been separated into two bodies, the Håland and the Helleren anorthosites, which show cross-cutting relationships and contrasting deformation features.

Based on intrusive relationships and structural features, the following units and relative chronology can be distinguished in the Rogaland Anorthosite Province. The *Egersund-Ogna Anorthosite* in the west forms the oldest body, with a foliated marginal zone that conforms to the layering in the surrounding Sveconorwegian gneisses. The body is separated from the similarly foliated *Håland Anorthosite* by a septum of highly strained, well-banded, Sveconorwegian gneisses. The relatively younger and undeformed *Helleren Anorthosite* in the central part of the province intrudes both the Egersund-Ogna and the

Håland bodies. The Helleren Anorthosite body may form the continuation of the likewise undeformed *Åna-Sira Anorthosite* in the eastern part of the province with a lobe of younger magmatic rocks (the Bjerkreim-Sokndal Layered Intrusion and jotunitic rocks) in between. The *Bjerkreim-Sokndal Layered Intrusion* is lying in an irregular trough structure on top of both the anorthosite bodies and the Sveconorwegian gneisses in the northeastern part of the province. The intrusion is capped by *mangerite and quartz mangerite,* which extends southeastwards as the so-called *Apophysis Intrusion* along the eastern margin of the Åna-Sira Anorthosite body. *Late-magmatic jotunitic to mangeritic rocks* occur as dykes and bodies in most of the Rogaland Anorthosite Province. The largest of these are the branched *Eia-Rekefjord jotunite intrusion* in the trough between the Helleren and Åna-Sira Anorthosite bodies, and the *Hadland jotunitic intrusion* at the western margin of the Egersund-Ogna Anorthosite body. Another group consists of *noritic intrusions* that form smaller bodies in the southeastern half of the province (map legend 16-19). Important among these is the *Tellnes body* in the Åna-Sira Anorthosite, which contains a world-class ilmenite deposit. Two smaller bodies of anorthosite and a granite body in the eastern part of the Rogaland Anorthosite Province lie isolated within the Sveconorwegian gneisses, and their ages relative to the other magmatic units, and to each other, thus cannot be determined from field relations alone. These bodies are the Garsaknatt Anorthosite east of Lundevatnet and the *Hidra Anorthosite* and *Farsund Charnockite* in the southeasternmost part of the province. The Rogaland Anorthosite Province was long after its consolidation transected by WNW-ESE-trending dykes. The 616 Ma old dolerite dykes form the *Egersund dyke swarm,* and were injected during the pre-Caledonian rifting just prior to the initial opening of the Iapetus Ocean (Bingen et al. 1998).

Though a sequential evolution can be demonstrated for most of the magmatic units in the Rogaland Anorthosite Province from the field relationships, age determinations have shown that the duration of magmatism in the province was about 10 million years, which is surprisingly short. U-Pb dating of zircon and baddeleyite extracted from orthopyroxene megacrysts (Schärer et al. 1996) yielded emplacement ages of 929± 2, 932 ± 3 and 932 ± 3 Ma for the Egersund-Ogna, Helleren and Åna-Sira Anorthosites, respectively, while late-magmatic injection of jotunitic dykes occurred at 931±5 Ma and the intrusion of the Tellnes noritic body at 920 ± 3 Ma.

The Egersund-Ogna Anorthosite

The Egersund-Ogna Anorthosite body is the largest in the Rogaland Anorthosite Province with a more or less dome-shaped, concentric, lithological structure (Maquil 1980 (unpublished map reproduced in Duchesne & Maquil (1987)), new mapping by NGU). The outer part of the body shows a variably developed foliation which follows the concentric structure, and which is concordant with the foliation in the Sveconorwegian gneiss envelope. In the south, this gneiss envelope continues as a thin septum of highly strained, sheared gneisses that separate the Egersund-Ogna Anorthosite from the Håland Anorthosite. The anorthosites at both sides of the septum show a penetrative planar foliation that is best developed on the northern side. The foliated marginal zone of the body has been interpreted as having formed during diapiric rise of the anorthosite as a crystal mush from mid- and lower crustal levels (Barnichon et al. 1999).

The central part of the Egersund-Ogna Anorthosite body consists of medium-grained homogeneous anorthosite that, in its centre northwest of Hellvik, contains phenocrysts and megacrysts of orthopyroxene and plagioclase. The central part of the body is also the part where labradorising plagioclase, both in matrix and phenocrysts, occurs most extensively in the province (Heldal & Lund 1995). Outwards follows a 1-4 km-wide zone of medium-grained leuconorite, which again is followed by homogeneous anorthosite that locally may contain scattered orthopyroxene and smaller, <5-10 cm-long, commonly rounded phenocrysts of plagioclase. On Nordre Eigerøya (in the west), this zone includes an area that is intruded by rather coarse-grained leuconorite forming an agmatitic structure with, on average, 20% leuconorite. Then follows a 0.5-1.2 km-wide, strongly foliated zone of anorthosite and leuconorite with 2-10 cm-

long, drawn out streaks and 10-30 cm-long, fish-like phenocrysts of orthopyroxene forming a well-developed planar structure. In places, trails of elongated larger orthopyroxene pheneocrysts also occur. The foliated zone has, in the north, sharp boundaries to little strained anorthositic units on each side, but transgresses to the margin of the Egersund-Ogna body in the eastern and southern parts. The two zones, which occur in the northern marginal part of the body, wedge out west of Helleland in the east and northwest of Bjårvatnet in the west. Immediately north of the foliated zone there is a zone of quite massive anorthosite with a varying density of generally <30-40 cm-long, orthopyroxene megacrysts of irregular shape. In places the massif anorthosite contains spectacular orthopyroxene megacrysts or clusters of megacrysts that are up to 1-2 m in length. The outermost zone in the north consists of homogeneous, medium-grained leuconoritic anorthosite with 5-10% of evenly distributed orthopyroxene. Locally, it contains scattered, <5 cm-long, rounded phenocryst of plagioclase. As in the orthopyroxene megacrystic zone, a foliated structure is weakly developed or absent, except in a narrow zone at the western margin, 3-8 km north of Bjårvatnet.

A jotunitic intrusion, *the Hadland Intrusion,* is located in the northwesternmost part of the Rogaland Anorthosite Province along the contact between Sveconorwegian granulite-facies gneisses and the Egersund-Ogna Anorthosite body. The intrusion consists essentially of jotunite and is mostly modally layered with a layering striking parallel to the contacts and dipping to the northwest.

The Håland Anorthosite

The Håland Anorthosite body lies to the south of the Egersund-Ogna Anorthosite, and is separated from this body by a strongly foliated contact zone that dips steeply to the north. The contact zone includes a septum of similarly highly strained Sveconorwegian gneisses, which separates the two bodies. In the east, the Håland Anorthosite is intruded and cut by the undeformed Helleren Anorthosite. The Håland Anorthosite is composed of anorthosite and leuconorite with complicated relationships between the different

rock phases. These have, in addition, suffered varying degrees of deformation, thus making the relationships even more complex.

The Håland Anorthosite body can basically be divided into two units. The northern unit is dominated by anorthosite with pseudo-enclaves of leuconorite (Michot 1961), which show varying degrees of deformation. The least strained, leuconorite pseudo-enclaves, as at Eigerøya, are 10-40 cm long and bun-shaped with elliptical cross sections. With increasing deformation, the pseudo-enclaves are stretched and thinned and eventually form a banded anorthositic rock with, discontinuous, dark orthopyroxene-bearing bands, up to some centimetres thick, which is often seen in the eastern part of the body. Along the northern margin of the body, immediately south of the gneiss septum from Koldal to Kydlandsvatnet, there is a 100 m-thick zone of seemingly undeformed leuconorite, which appears to host most of the ilmenite ore deposits, which are found in this part of the province. The leuconorite zone widens in the east to become 300-400 m thick north of Kydlandsvatnet.

The southern unit of the Håland Anorthosite consists of interlayered anorthosite and leuconorite with enclaves of phases showing modal-like layering. The dominant rock type is a coarsely spotted leuconorite with 2-4 cm-long, elongated orthopyroxene clusters that define a distinct, planar, foliated structure. The spotted leuconorite has local, 0.5-2 m-thick, foliation-parallel layers of anorthosite in addition to plentiful diffuse veins of anorthosite (several generations?), which seem to intrude all other phases. Quite commonly there are fish-like enclaves, up to several metres in length, of a modally banded phase with close-lying mafic layers, up to a few cm thick. This phase looks much like the high-strain banded anorthositic rock in the northern unit of the body and may represent inclusions of this unit in the spotted leuconorite. The foliated structure of the spotted leuconorite is folded into tight folds in the eastern part of the body, where it is cut by the undeformed Helleren Anorthosite. It thus appears likely that the southern part of the Håland Anorthosite was, for the most part, intruded and deformed after the emplacement

and deformation of the northern unit.

The Helleren Anorthosite

The Helleren Anorthosite body cuts across the boundaries and internal deformation structure of both the Egersund-Ogna and the Håland Anorthosite bodies. It has clear intrusive contacts which, in the north and east, are transected by the Bjerkreim-Sokndal Layered Intrusion. The Helleren Anorthosite consists of massive, rather coarse- to medium-grained anorthosite and leuconorite, which show little or no signs of structural deformation. Leuconorite occurs first of all in a 1-2 km-wide zone at the western margin of the body, but patches of leuconorite also occur in the anorthosite farther east. The leuconorite is usually medium-grained with evenly dispersed ortopyroxene (spotted). Finer-grained varieties occur in the west-facing embayment into the Håland Anorthosite (Michot 1961). The anorthosite may locally contain phenocrysts of plagioclase while orthopyroxene phenocrysts are relatively rare.

The Åna-Sira Anorthosite

The Åna-Sira Anorthosite is a fairly homogeneous anorthosite massif which consists essentially of anorthosite and anorthositic leuconorite (Krause et al. 1985). These rock types are irregularly interspersed and have not been differentiated on the map. The composition and texture of the anorthosite appear to be similar to those of the Helleren Anorthosite. Both the Åna-Sira and the Helleren anorthosites have been dated to 932 ± 3 Ma by U-Pb isotopes (Schärer et al. 1996). Plagioclase chemistry reveals a NW-SE elongated pattern with the most primitive compositions in the central part of the anorthosite body (Zeino-Mahmalat & Krause 1976). To some extent, a similar NW-SE trending pattern is displayed by mineral lamination and sporadic modal layering (Knorn & Krause 1977). The Åna-Sira Anorthosite is the most significant anorthosite massif in Rogaland in terms of Fe-Ti oxide ore deposits and hosts two giant ilmenite ore bodies, the *Tellnes* and *Storgangen* deposits, as well as a number of smaller deposits. The anorthosite is cross-cut by a several km-long norite pegmatite (the *Blåfjell-Laksedal pegmatite*) which is associated with a suite

of coarse-grained Fe-Ti oxide mineralisations. The *Hogstad Intrusion* is located in the southeastern part of the Åna-Sira Anorthosite and consists of an eastern leuconoritic portion and a western noritic portion. It displays sub-vertical layering and a cumulate zoning equivalent to parts of the Bjerkreim-Sokndal Layered Intrusion (Vander Auwera & Duchesne 1996). The Åna-Sira Anorthosite also contains a small, layered noritic body (the *Bøstølen Intrusion*) as well as a number of dykes, ranging in composition from jotunite to norite and mangerite. Megacrystic Ca-poor pyroxenes occur only sporadically in the Åna-Sira anorthosite.

The Bjerkreim-Sokndal Layered Intrusion

The Bjerkreim-Sokndal Layered Intrusion consists of a layered cumulate sequence ranging in composition from anorthosite and troctolite to norite, gabbronorite, jotunite and mangerite. The intrusion comprises a Layered Series, more than 7000 m thick, suggested to be derived by fractional crystallization of jotunitic parental magmas (Duchesne & Hertogen 1988, Robins et al. 1997). It is capped by mangeritic, quartz-mangeritic and charnockitic rocks. The Layered Series may be divided into 6 megacyclic units (MCUs), each resulting from magma replenishment episodes, and within each MCU a characteristic *cumulate* sequence is partly or fully repeated. With differentiation of the magma, the cumulus minerals crystallized in the general order: plagioclase - ilmenite (or plagioclase + olivine + ilmenite + magnetite) - Ca-poor pyroxene (olivine and magnetite out) - magnetite - Ca-rich pyroxene and apatite - fayalitic olivine, but with minor variations (Wilson et al. 1996). On the map the cumulate nomenclature of Wilson et al. (1996) has been maintained in order to provide as detailed an account as possible of the intrusion petrology.

The intrusion is traditionally divided into three exposed lobes; the Bjerkreim, Sokndal and Mydland lobes (Michot 1960), although the Mydland lobe has been shown to be an enclave within the jotunitic-mangeritic *Apophysis Intrusion* (Bolle et al. 1997). The other lobes have been deformed into a roughly N-S elongated syncline or trough, due to gravitational subsidence of the core of the intrusion

at a late magmatic stage (Paludan et al. 1994, Bolle et al. 2002). The cumulate sequence is thickest along the south-plunging axis of the northern Bjerkreim lobe, where the most complete sequence of MCUs is found. The acidic rocks, which cap the Layered Series, comprise quartz mangerites and charnockites (Duchesne & Wilmart 1997), which seem to be continuous with the Apophysis Intrusion (Rietmeijer 1979, Bolle et al. 1997) extending along the northern and eastern contacts of the Åna-Sira Anorthosite.

The Apophysis Intrusion

The Apophysis Intrusion (Bolle 1996) extends from the quartz-mangeritic cap of the Bjerkreim-Sokndal Layered Intrusion towards the southeast along the contact between Sveconorwegian gneisses and the Åna-Sira Anorthosite. The intrusion is composed of diverse, intermingled, jotunitic to quartz-mangeritic rock types.

The Hidra and Garsaknatt anorthositic-leuconoritic intrusions

Hidra and Garsaknatt are two smaller, dominantly anorthositic to leuconoritic bodies, intruded into the gneissic envelope (Michot & Michot 1969). The Hidra body contains a marginal chill zone that provides important evidence for an origin by fractional crystallization of a jotunitic parental magma (Demaiffe & Hertogen 1981). In addition, the body is cross-cut by a stockwork of charnockitic dykes (Weis & Demaiffe 1983). The Garsaknatt body shows clear intrusive relationships with the host Sveconorwegian gneisses and comprises anorthosite and leuconorite as well as segments of modally layered norite. Both the Garsaknatt and the Hidra bodies contain resources of iridescent plagioclase-bearing anorthosite and are potential candidates for dimension stone extraction (Heldal & Lund 1995).

The Eia-Rekefjord Intrusion and jotunite dykes

The Eia-Rekefjord Intrusion (Michot 1960) is the most prominent jotunitic intrusive body in the province. It is found primarily along the contact between the southern lobe of the Bjerkreim-Sokndal Layered Intrusion (the Sokndal lobe) and the neighbouring anorthosite massifs, but also transects the quartz mangerites capping the intrusion. The Eia-Rekefjord Intrusion intruded the Sokndal lobe cumulates and the Åna-Sira Anorthosite, and related jotunitic rocks are most probably found as a marginal intrusion in the Mydland lobe. The Eia-Rekefjord Intrusion generally consists of rather monotonous medium grained jotunite.

Other jotunitic intrusive rocks are present throughout the province as dykes cross-cutting all the major anorthosites (Michot 1960). The most significant jotunite dyke is the Lomland dyke, which originates in a norite body in the Håland Anorthosite, cuts the Håland and Egersund-Ogna anorthosites and forms a complex branching pattern upon entering the Bjerkreim lobe of the Bjerkreim-Sokndal Layered Intrusion. The compositions of most of the jotunite intrusive rocks are controlled by a well defined liquid line of descent, giving rise to a continuous compositional range from primitive jotunites to quartz mangerites (Vander Auwera et al. 1998).

The Farsund Charnockite

The Farsund Charnockite is one of the acidic intrusions associated with the Rogaland Anorthosite Province. It is a dark olive-green rock dated by U-Pb isotopes to between 900 and 940 Ma (Falkum & Petersen 1987), with clear potassic affinities and characterised by distinctly higher Fe/Mg ratios than the hornblende-biotite granitoids elsewhere in South Norway.

Sveconorwegian gneisses

The granulite-facies Sveconorwegian gneiss complex in Southwest Rogaland (Hermans et al. 1975) constitutes the host rocks of the Rogaland Anorthosite Province and is exposed in the northeast of the province and in a septum between the Egersund-Ogna and Håland anorthosites. Individual gneiss units generally show large lateral continuity which may be partly a primary intrusive feature but is also due, not least, is due to attenuation during the deformation. The gneisses are predominantly of igneous origin and can be broadly divided into two groups. The older group consists of granodioritic to granitic orthogneisses showing partial-melt veins while the younger group con-

sists of various types of granites without anatectic veining. Both are deformed and foliated. The gneisses of the older group in particular show strong deformation features including, quite commonly, a marked planar fabric in which primary relations between the different rock components are more or less obliterated and melt veins have formed parallel to the foliation. These well-foliated gneisses are usually referred to as *banded gneisses* despite often showing a banding composed only of stretched, parallel melt veins. In the map area, the banded gneisses include a few minor layers or bands of highly anatectically melted, garnet-sillimanite-bearing *pelite* and, less frequently, orthopyroxene-bearing *amphibolite*. The pelitic gneiss commonly contains ossumilite that formed during contact metamorphism. Thin *quartzite* bands are only important in one place, near Gjersdal, in the southeastern part of the area. The last phase of the banded gneisses is a *light grey metagranite* which occurs as deformed, almost conformable veins or bands of varying density. The rocks and high degree of strain in the gneiss septum at the Egersund-Ogna-Håland Anorthosite boundary are very similar to those in the banded gneisses.

The younger group of granitic gneisses without partial-melt veins comprises even-grained and porphyritic types. More or less foliated, medium-grained *granitic gneiss* is the most common even-grained type. A rather coarse-grained *metagranite rich in white pegmatites* occurs near Hovsvatnet. The porphyritic types are weakly to well foliated. Most common, and particularly westwards from Hovsvatnet, is a *porhyritic metagranite* with a varying density of deformed K-feldspar phenocrysts. West of Lundevatnet occurs a characteristic coarse-porphyritic metagranite belonging to the Feda suite (Bingen & van Breemen 1998a, 1998b). Both types of porphyritic granite contain variable amounts of scattered, deformed veins of metagranite.

The degree of deformation is generally high in the Sveconorwegian gneisses with the highest strains recorded in the banded gneisses, which may have evolved in a ductile shear regime at a deep crustal level. As is common in granulite-facies terrains, lineations are poorly developed, but when observed they mostly plunge gently to the east or west. An exception is the highly strained gneisses at the contact to the Apophysis Intrusion where lineations are sub-vertical and quite common. Also, the mangeritic rocks at the eastern margin of the Apophysis Intrusion show a distinct foliation.

The age of the Sveconorwegian gneisses is not fully resolved, particularly for the banded gneisses. The intrusion of the porphyritic granite of the Feda suite has been dated at 1051 Ma (Bingen & van Breemen 1998a), while granitic gneisses give magmatic ages around 1035-1056 Ma (Möller et al. 2002). The age of peak metamorphism, as determined from U-Pb monazite dating (Bingen & van Breemen 1998b), is around 1025-970 Ma, and thus at least 40 million years older than the intrusion of the anorthosite plutons. Möller et al. (2002) obtained U-Pb ages for metamorphic zircons at c. 1015-1000 to c. 910-930 Ma in granitic gneisses north of Teksevatnet.

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Mineral resources in the Rogaland Anorthosite Province, South Norway: origins, history and recent developments

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The Neoproterozoic Rogaland Anorthosite Province (RAP) in Southwest Norway contains significant mineral resources. Most important are the ilmenite ores, of which the world-class Tellnes orebody is currently in production. Tellnes supplies some 7% of the total world titanium consumption. During the last decades ilmenite refiners have been increasingly demanding with respect to trace elements, which have widened the targets for ilmenite exploration in the area. A review of resources, quality and the current state of knowledge on their genesis is presented.

Sulphide mining in the RAP was abandoned by the end of the 19th century, but 11 occurrences of Fe-Ni-Cu sulphides are known. Only one ore body, the Homsevatnet deposit, is of significant size, but all occurrences are similar in terms of paragenesis and composition. All known deposits are poor in noble metals, implying that little interaction took place between immiscible silicate and sulphide melts. In addition, most mafic intrusions contain subsidiary amounts of dispersed sulphide droplets, but only in the Bjerkreim-Sokndal layered intrusion did a segregation process lead to increased grade and tenor.

Current resources also include dimension stones. Anorthosite containing blue-iridescent plagioclase (*Labrador Antique*) is at the moment the most valuable dimension stone produced in Norway. Ongoing exploration is demonstrating the existence of large quantities of exploitable dimension stone resources. Aggregate is extracted at four quarries from white altered anorthosite as well as from jotunite, and historically, kaolin has been mined on a limited scale.

The identified resources of ilmenite (± apatite, ± vanadian magnetite), dimension stone and aggregate are all very large. In terms of current extraction, exploration and future prospects, the RAP ranks as one of the most important mineral provinces in Norway.

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Introduction

The Neoproterozoic massif-type anorthosite province in Rogaland county, South Norway, has been the target of mining and exploration for more than 200 years. The province is currently being exploited for a wide selection of mineral commodities. The economically most important product is ilmenite. The province contains a variety of ilmenite deposits, the best known being the Tellnes, Storgangen and Blåfjell deposits. Since 1965, Tellnes has been the only ilmenite mine in production in the province and presently the deposit supplies around 7% of the world's titanium consumption. More than 80 mines and prospects are known (Fig. 1).

Changing processes in the production of TiO_2 as well as environmental concerns have focussed attention on the chemical specifications of titanium raw materials. For ilmenite, magnesium is the main issue, because magnesium basically controls which recovery processes can be applied to a given ilmenite feed stock. In TiO_2-pigment production, a chloratable feed stock must contain less than 1.5% MgO (Korneliussen et al. 2000). Other deleterious elements in pigment production include Cr and Ni, which may reside in ilmenite, and pollutants such as Ca, which mainly derive from silicate minerals in the concentrates. In Rogaland, much effort has recently been expended to locate potential deposits of low-MgO ilmenite in conjunction with vanadian magnetite and apatite. Several noritic rock bodies contain in excess of 10% apatite, and significant resources with 30-35% ilmenite + apatite + vanadian magnetite have been identified.

Several semi-massive to massive sulphide deposits located within the anorthosite bodies have been mined or explored for Ni and Cu. The largest is the Homsevatnet deposit, located 12 km northeast of Egersund (Fig. 1), where mining took place during the 1870s. Several noritic intrusions in Rogaland contain dispersed base metal sulphides in amounts of around 1%. In the Bjerkreim-Sokndal Layered Intrusion there is a stratiform sulphide occurrence with sulphide contents of up to around 10%. Disseminated sulphides are recovered as a by-product of ilmenite exploitation at Tellnes, primarily for their nickel content. Today, the economic potential of the Rogaland sulphide deposits seems to be insignificant.

Anorthosite, in different varieties, and associated jotunitic rocks are quarried for aggregate, and the presence of iridescent plagioclase in the anorthosites have recently given rise to a dimension stone industry. Anorthosite from Rogaland containing blue-iridescent plagioclase is sold under the name *'Labrador Antique'*. Mining for kaolin has taken place at several localities in the anorthosite bodies, but was abandoned at the beginning of the 20th century.

The purpose of this paper is to review the history, occurrence and potential of the Rogaland mineral deposits and summarize current ideas on their formation.

Geological setting

The RAP (Fig. 1) is located along the coast of southwestern Norway, roughly between Ognabukta in the north and Lindesnes in the south. It covers an onshore area of about 1750 km², and a similar area offshore (Sigmond 1992). The province is dominated by four major massif-type anorthosite plutons; the Egersund-Ogna, Håland, Helleren and Åna-Sira anorthosites, which are also the oldest intrusive units in the province. The anorthosites consist mostly of monotonous anorthosite or leuconorite. The Egersund-Ogna massif contains a marginal zone of foliated leuconorites, presumably formed through syn-emplacement deformation (Duchesne & Michot 1987), and the Håland massif mainly comprises foliated leuconorites. Metre-sized aggregates of megacrystic aluminous and Cr-rich orthopyroxene are found to variable extent in all massifs. They bear evidence for a polybaric origin of the anorthosites and indicate initial pressures of crystallization of 11-13 kbar (Longhi et al. 1999, Vander Auwera et al. 2000). Final emplacement of the anorthosites and associated intrusions probably took place at around 5 kbar (Vander Auwera & Longhi 1994). The whole province was formed during a brief period of time bracketed between the emplacement of the major anorthosites at 932-929 Ma and the emplacement of the Tellnes ilmenite deposit in the Åna-Sira anorthosite at 920 ± 3 Ma (Schärer et al. 1996).
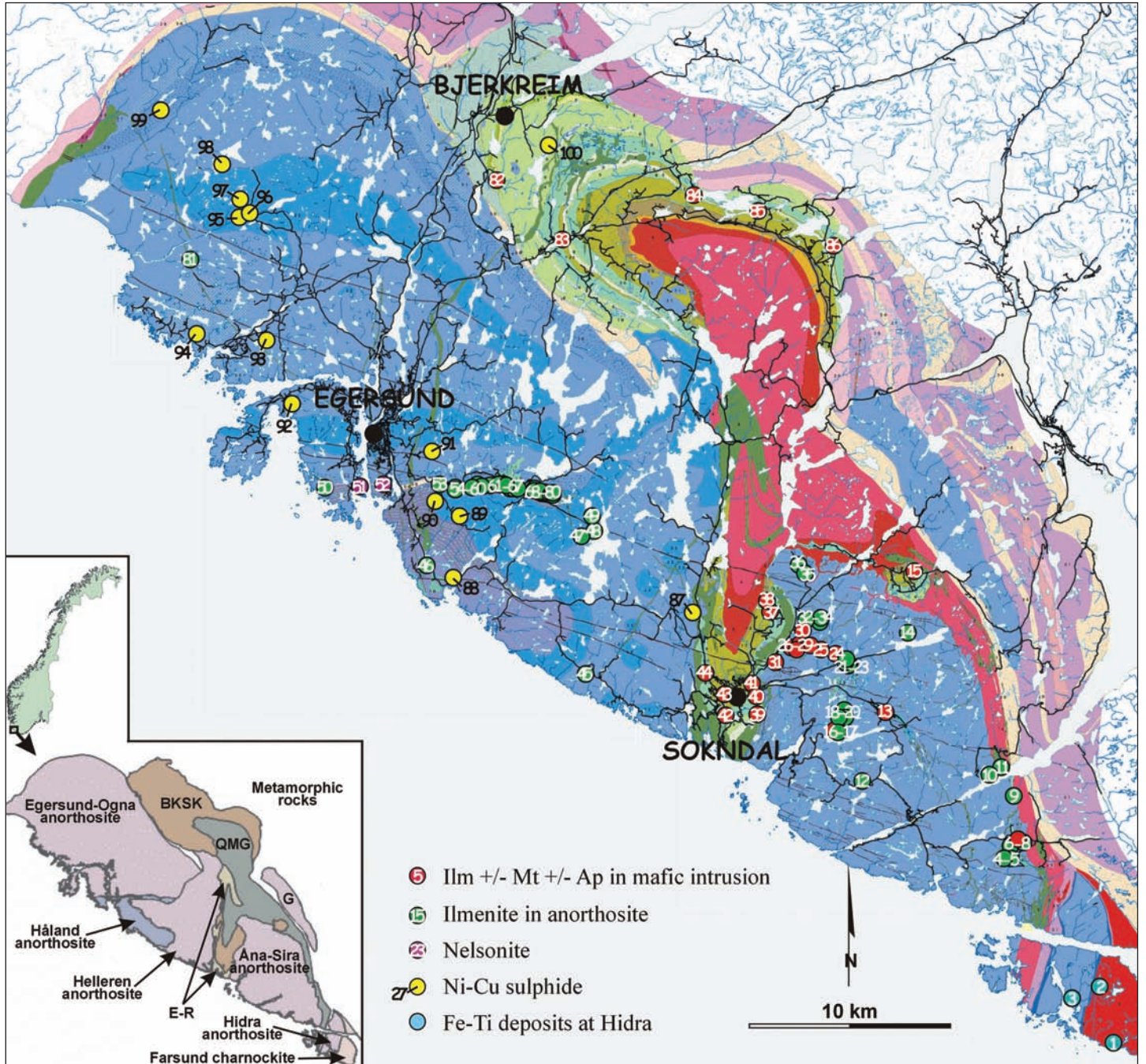
Fig. 1. Map of the western part of the Rogaland Anorthosite Province (Marker et al. 2003) showing registered prospects and deposits of ilmenite (±magnetite, ±apatite) and Fe-Ni-Cu sulphides. Anorthositic rocks are shown in blue colours, jotunitic and noritic rocks in yellow-green and granitoid rocks in red (see Marker et al. (2003) full map legend). Inset map shows the major intrusive units; BKSK: Bjerkreim-Sokndal Layered Intrusion, QMG: quartz-mangeritic rocks, E-R: Eia-Rekefjord Intrusion, G: Garsaknatt Intrusion. Numbers refer to deposits. Ilmenite deposits: 1: Saksekjær, 2: Veisdal, 3: Tinnbakkneset, 4: Vardåsen, 5: Hilleråsen, 6: Grasjenåsen, 7: Skolla skjerp, 8: Kvannvikåsen, 9: Dalestøa skjerp, 10: Strandbakken, 11: Gjersdal, 12: Raunslid, 13: Tellnes, 14: Brombo, 15: Mydland, 16: Bøstølen, 17: Sletthei, 18: Laksedal, 19: Dalens gang, 20: Bryns skjerp, 21: Blåfjelldalen, 22: Blåfjell (several adits), 23: Li gruve, 24: Aursland skjerp, 25: Brekke gruve, 26-29: Storgangen gruver, 30: Sidegangen, 31: Åmodt gruve, 32: Kjørfjell, 33: Blåfjellskaret, 34: Frøytlog, 35: Ålgård, 36: Florklev, 37: Bakka, 38: Ørsland, 39: Årstad gruve, 40: Prestbro skjerp, 41: Bø utmark, 42: Årstadøy, 43: Hauge gruve, 44: Drageland, 45: Vatland, 46: Svånes, 47: Pramstø, 48: Pramsknuten/Jerneld, 49: Snøringdalen, 50: Vaksviken, 51: Eigerøy, 52: Hestnes, 53: Tysslandsvad, 54: Rødemyr gruve, 55: Fisketjern gruve, 56: Berkjedalen, 57: Thorsdalen gruve, 58: Skåra gruve, 59: Raaes gruve, 60: Kaknuten gruve, 61: Bredeskaret, 62: Sørskaret, 63: Omdalsskaret, 64: Vallbakken, 65: Langemyr, 66: Karen gruve, 67: Knuts gruve, 68: Lyngnes gruve, 69: Lundetangen, 70: Peder Anker, 71: Ankershus, 72: Egebakkskaret (W), 73: Egebakkskaret (E), 74-75: Lielven gruver, 76-77: Furufjellet gruver, 78: Jentofts gruve, 79-80: Sandknuten gruver, 81: Beinskinnvatnet, 82: Åsen, 83: Helleland, 84: Teksevatnet, 85: Tekse-Bilstad, 86: Vasshus. Sulphide deposits: 87: Urdal, 88: Grastveit, 89: Koltjørn, 90: Helland, 91: Skora, 92: Myklebust, 93: Hellvik, 94: Piggstein, 95-96: Uland, 97: Bjørndalsnipa, 98: Homsevatnet, 99: Fossfjellet, 100: Rudleknuten.

A number of slightly younger intrusions are found in close spatial association with the main anorthosite bodies. These include two smaller anorthositic/leuconoritic bodies (the Hidra and Garsaknatt intrusions), the 230 km² Bjerkreim-Sokndal Layered Intrusion and the Eia-Rekefjord jotunitic intrusion (Fig. 1). In addition, the province is cross-cut by numerous noritic, jotunitic and mangeritic dykes of variable width and lateral persistence.

Acidic rocks in the province are represented by quartz mangerites and charnockites which cap the Bjerkreim-Sokndal Intrusion. Their status has been interpreted as contaminated differentiates of the Bjerkreim-Sokndal Layered Intrusion (Nielsen et al. 1996), possibly mingled with externally-derived acidic magma (Duchesne & Wilmart 1997). The Farsund Charnockite in the eastern part of the province (Fig. 1) is also considered a part of the anorthosite province (Duchesne et al. 1985).

Ilmenite

Mining history

The existence of significant Fe-Ti oxide deposits in Rogaland was recognised more than 200 years ago, when a number of iron mines were opened in ilmenite deposits around Lake Kydlandsvatnet in the western part of the province in 1785 (Krause et al. 1985). However, Moss Jernverk, who operated the mines, never managed to smelt the Ti-rich ore and closed down operations after only a few years of activity. During the following century, exploration located Fe-Ti oxide deposits at Storgangen, Blåfjell and Flordalen (Dahll 1863), and in the mid-1800s mining was re-started here and at several localities in the southern (Sokndal) lobe of the Bjerkreim-Sokndal Layered Intrusion (Carlson 1945), as well as in the Kydlandsvatn area. However, the high titanium content again limited the value of the deposits as iron ore, and by 1881 industrial-scale production had ceased in the province.

In 1910, JHL Vogt, after decades of dedicated research on the Rogaland Fe-Ti prospects, regrettably concluded that the Rogaland deposits would have little economic value until further use for titanium had been found (Vogt 1910). Only a few years later, however, pioneering research by G. Jebsen and P. Farup resulted in the development of a new industrial method for producing a white TiO_2 pigment (now called the *sulphate process*), and demand for titanium minerals dramatically increased. By 1917, Titan Co. A/S, Frederikstad, had acquired the majority of the shares in the mining company Titania A/S (founded in 1902), that owned the rights to the Rogaland deposits, in order to secure the raw materials for their new titanium pigment plant.

A beneficiation plant was constructed at Sandbekk near the Storgangen orebody, then the largest known prospect in the province, and industrial-scale mining commenced in 1916. Reserves at Storgangen were estimated to at least 200,000t of ore – enough for 10 years of production. During the following two decades Storgangen was one of the largest titanium mineral producers in the world, and when mining finally ceased in 1965 the deposit had yielded more than 10 Mt of ore. More than 60 Mt of reserves remain (Hagen 1992).

By 1965, all production had shifted to an open-pit mine at Tellnes also within the Åna-Sira anorthosite body. Here, three km southeast of Storgangen, the existence of an elongated noritic lens had been known since the late 19th century (Kolderup 1896), but its potential as an ilmenite resource was only recognized following aeromagnetic surveys in the mid-1950s. Exploitation continues at Tellnes today – the Tellnes ore reserve of 380 Mt with more than 60 Mt contained TiO_2 represents around 15% of the world's titanium mineral reserves, and production at Tellnes currently accounts for 7% of the total global production of titanium minerals.

Some of the ilmenite deposits and ilmenite-rich rocks in Rogaland carry significant amounts of vanadian magnetite and apatite. Recently, attention has been directed towards the large, but low-grade, gabbronorite-hosted, ilmenite occurrences in which ilmenite may be exploited in conjunction with apatite and vanadian magnetite.

Occurrence and genesis

The RAP contains a wide range of Fe-Ti deposits, from small bodies of massive ilmenite to larger volumes of modally layered rocks in layered intrusions. A cumu-

late origin is usually accepted for ilmenite-rich rocks found as concordant layers or zones in layered intrusions, such as in the Bjerkreim-Sokndal (Fig. 1), Hogstad (Fig. 1, #7), Storgangen (#26-29) and Bøstølen (#16) intrusions. However, within the anorthosite massifs oxide-rich rock commonly occurs in minor bodies of more or less homogeneous composition, usually characterized by a simple mineralogy. In these cases the origin is much more obscure. Nevertheless, a continuous range of compositions does exist, from pure oxide rocks to oxide-plagioclase rocks and oxide-rich norites grading into multiphase assemblages which are generally layered (Table 1). Duchesne (1999) inferred a cumulate origin for all non-nelsonitic deposits in the region.

'Primitive' ilmenite-rich intrusions of more or less mono- or bi-mineralic composition are widespread in Rogaland. Modal layering in these rocks is either rare or rudimentary, and the intrusions are found mainly as irregular dykes or lenticular bodies. Most of these primitive ilmenite-rich rocks consist of masses of almost pure ilmenite, some covering an area of up to a few 100 m². Texturally, the ilmenite is thoroughly equilibrated, forming coarse-grained polygonal aggregates, with interstitial green spinel (pleonast) and locally subsidiary amounts of magnetite. However, isolated monomineralic ilmenite bodies are rare, and most primitive deposits are, in fact, spatially associated with noritic dykes or lenses of variable dimensions.

The Pramsknuten (Fig. 1, #48) (or Jerneld) deposit (Duchesne 1999) is one of a small number of deposits which do not seem to contain silicates at all. The deposit is composed of a string of small oxide lenses in an elongated zone approximately 200 m long within the Helleren anorthosite. The deposit is more or less confined to a zone of 'white' (altered) anorthosite. Like the other massive oxide bodies, the ilmenites are texturally equilibrated and a cumulate origin is speculative. Sulphides are abundant and consist predominantly of pyrite. As in all primitive ilmenite deposits ilmenite grains contain a relatively high amount of exsolved hematite (Duchesne 1999) and are more properly termed hemo-ilmenite. The ilmenite composition is characterized by an MgO content of around 5%, which is among the highest MgO contents in ilmenite recorded in Rogaland. The Cr_2O_3 content of ilmenite is generally positively correlated with MgO and exceeds 0.4% in the Pramsknuten ilmenite (Duchesne 1999, Duchesne & Schiellerup 2001).

Bi-mineralic oxide deposits contain plagioclase in addition to hemo-ilmenite and are found, for instance, at Florklev and Ålgård in the Åna-Sira Anorthosite (Fig. 1, #35,36). The mineralization here is related to a narrow dyke in which grains of little-deformed, tabular, and generally laminated plagioclase are dispersed in a recrystallized ilmenite matrix. In this case the presence of significant amounts of euhedral plagioclase (10-50%) indicates that the deposits have a cumulate origin, and that all subsequent deformation was accommodated by oxide recrystallization. In the dyke, modal variations delineate subtle layering. This type of occurrence qualify as plagioclase-ilmenite cumulates (piCs). Plagioclase-ilmenite cumulates also make up large parts of the Bjerkreim-Sokndal Layered Intrusion (Wilson et al. 1996, Robins & Wilson 2001). Generally, however, these cumulates are anorthosites or leuconorites with only a few percent ilmenite. Only when magma influx and mixing in the Bjerkreim-Sokndal magma chamber took place did more oxide-rich plagioclase-ilmenite (± Ca-poor pyroxene) cumulates form (Jensen et al. 1993) (see below).

Most massive oxide deposits are found in association with noritic rocks, characterized by the assemblage: hemo-ilmenite, plagioclase and orthopyroxene. The Blåfjell-Laksedal field (Fig. 1, #18-23) contains some of the most renowned deposits in the province, all of which are related to the same norite pegmatite (Krause et al. 1985). Seven adits at Blåfjell into the principal mineralizations of the pegmatite produced around 100,000 tons of ore during the second half of the 19th century. Though the exact relationship between the massive oxides and the norite pegmatite is indistinct, thick modal layers of monomineralic ilmenite in the pegmatite are developed in the vicinity of the Blåfjell mines. Local gradational contacts are observed between unmineralized norite and plagioclase-bearing massive ore. These observations seem to suggest a cumulate origin for the ilmenite ore bodies in the Blåfjell-Laksedal field.

In other classic deposits, such as Vardåsen, in the eastern part of the Åna-Sira Anorthosite (Fig. 1, #4), and at least locally in the Koldal field (#61-67), massive ilmenite ores are also spatially associated with norite dykes. Whereas the association with norite dykes in the Koldal field is unclear, the Vardåsen deposit displays a transitional contact in a north-south striking dyke, between a poorly mineralized layered noritic cumulate to the west and almost massive ilmenite to the east. In deposits where ilmenite is not concentrated into massive ore bodies but interspersed with plagioclase and orthopyroxene in noritic rock, the ilmenite ore is typically modally layered. This is the case in the Raunsli, Frøytlog and Svånes deposits (Fig. 1, #12,34,46).

In the Bjerkreim-Sokndal Layered Intrusion, plagioclase-ilmenite (± orthopyroxene) cumulates with up to 50% hemo-ilmenite are found at the base of two megacyclic units, where influx of fresh magma caused cryptic regressions in the Layered Series. Though these ilmenite-rich rocks have never been exploited they constitute a large titanium resource. The regressive sequence at the base of megacyclic unit IV (zone 'A' in Fig. 2) is up to 30 m thick and may be followed for more than 20 km along strike (Jensen et al. 1993). A single ilmenite analysis yielded 2.7% MgO, whereas ilmenite immediately above the zone contains 4.4% MgO and 690 ppm Cr_2O_3 (Duchesne 1999). The primitive basal oxide-rich cumulates alternate with anorthositic layers and crystallized from a hybrid magma, produced by mixing of evolved resident magma with inflowing primitive magma (Wilson et al. 1996).

In all the ilmenite deposits that contain ilmenite with or without cumulus plagioclase and orthopyroxene, the ilmenite contains extensive hematite exsolutions and is generally characterized by MgO contents of more than 4% and Cr2O3 contents above 4-500 ppm. Coexisting plagioclase has anorthite contents of 47-51% and orthopyroxene Mg-numbers between 73 and 78%.

Tellnes

The Tellnes deposit (Fig. 1, #13) is by far the largest deposit in Rogaland with reserves amounting to 380 MT with more than 60 MT contained TiO_2. Cross-cutting relationships, as well as U-Pb isotopic dating, imply that the emplacement of the Tellnes deposit is one of the youngest events in the Rogaland Anorthosite Province. The emplacement has been dated by the U-Pb method to 920 ± 3 Ma (Schärer et al. 1996).

Tellnes is a large, sickle-shaped, Ti-rich noritic ore-body almost three km long and more than 400 m wide. The intrusion is fairly homogeneous and modal layering is rare. The average modal composition of the Tellnes ore body is 53% plagioclase (An_{45-42}), 29% hemo-ilmenite, 10% orthopyroxene (En_{77-75}) and 4% biotite (Gierth & Krause 1973, Krause et al. 1985, Duchesne 1999). The remaining ~4% is made up by the accessory phases: olivine (which is commonly replaced by orthopyroxene-oxide symplectites), magnetite, Ca-rich pyroxene, apatite and sulphides. Though mineral lamination does occur, cumulate textures are not obvious. The mineralogy is essentially constant with more than 90% of any sample made up by the three phases plagioclase, orthopyroxene and ilmenite. However, the modal proportions of the minerals may vary considerably which results in a strong chemical zoning across the orebody (Robinson et al., this volume). The content of TiO_2 varies between 10 and 25% with some samples containing up to 35% TiO_2.

The texture is to a large extent dominated by subhedral to euhedral plagioclase laths. Ca-rich pyroxene is often present only as rims on orthopyroxene grains. The Tellnes deposit may therefore most appropriately be considered a three-phase cumulate of plagioclase, hemo-ilmenite and orthopyroxene (pihC), with the local presence of cumulus olivine. The minor phases are entirely of intercumulus origin.

Hemo-ilmenites from the central part of the orebody contain around 15% hematite component. They typically have MgO contents in excess of 3-4% and Cr_2O_3 contents of 400-550 ppm (Table 1, Krause et al. 1985, Duchesne 1999, Kullerud 1994). The MgO, Cr_2O_3 and hematite contents of ilmenite in the central part of Tellnes are thus typical for Rogaland deposit types with few coexisting mineral phases, such as the Raunsli, Frøytlog, Flordalen and Svånes deposits mentioned above. The
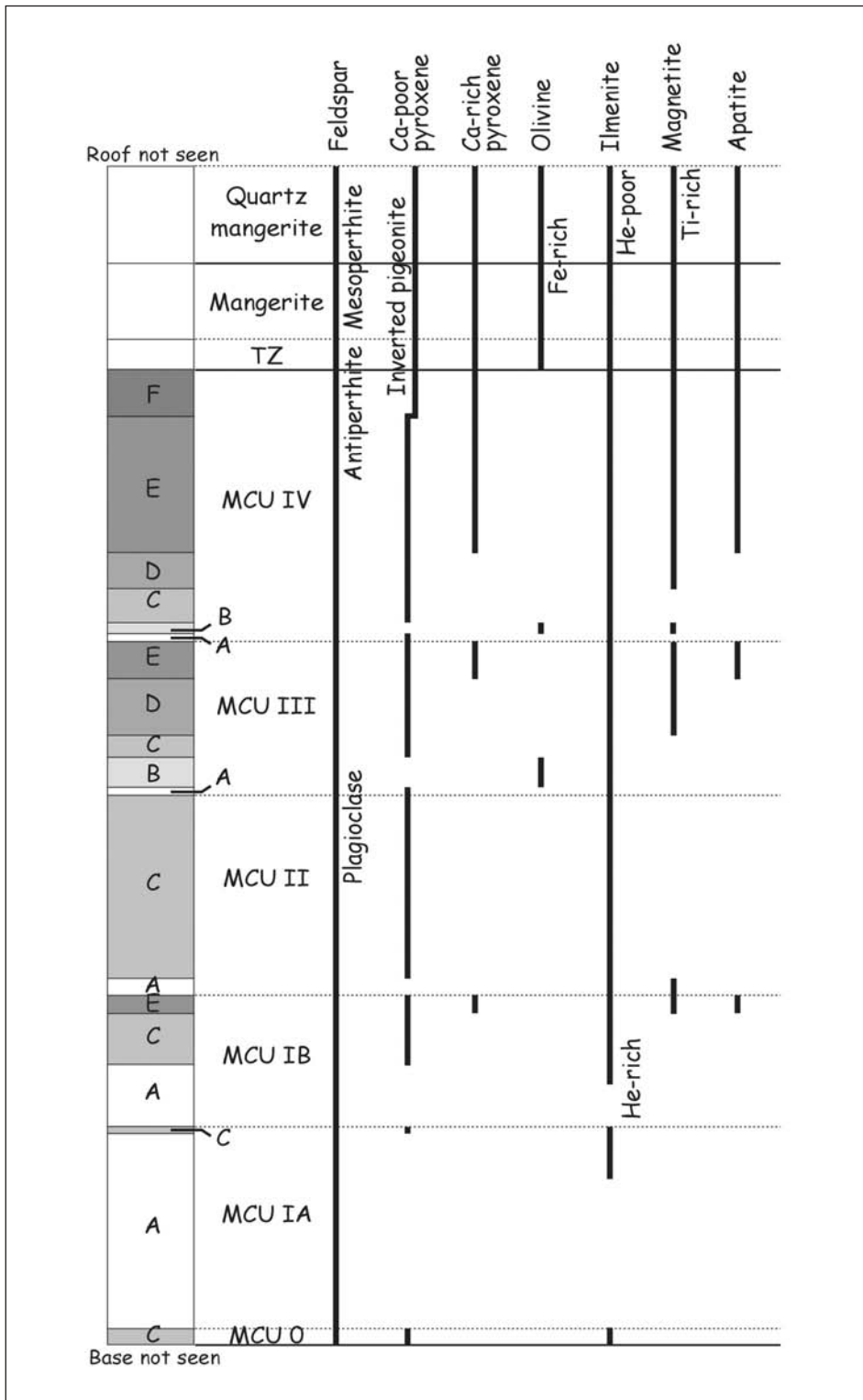
Fig. 2. Generalized stratigraphy and zone sequences in the Bjerkreim-Sokndal Layered Intrusion according to Wilson et al. (1996). The full sequence of MCUs, as shown here, is developed only in the (northern) Bjerkreim-lobe of the intrusion. The total thickness of the Layered Series is close to 7000 m. 'TZ': Transition Zone.

mush wet by an interstitial (Ti-rich) jotunitic melt (Wilmart et al. 1989). Recent experimental results have now confirmed that the late-stage melt must have been jotunitic in composition, but suggest that it was relatively Ti-poor (Skjerlie et al., this volume). Based on field relations, the Tellnes deposit was also thought to be genetically related to an associated mangeritic dyke, resulting in complex genetic models (Krause & Pedall 1980, Wilmart et al. 1989). However, U-Pb dating has revealed that the two intrusive units are separated in time by more than 10 Ma, rendering a genetic link improbable (Schärer et al. 1996). Similarly, a suggested relationship between the Tellnes orebody and the 10 Ma older Åna-Sira anorthosite, based on strontium and oxygen isotopes (Wilmart et al. 1989, 1994), also have to be abandoned. Robinson and coworkers (in this volume) suggest that mixing between a relatively evolved magma and a more primitive magma may have been important in the evolution of the chemical and modal composition of the Tellnes ore body.

Storgangen

The Storgangen (Norwegian for 'large dyke') Intrusion is a relatively thin modally-layered sheet emplaced into the Åna-Sira anorthosite (Fig. 1, #26-29). Storgangen is the second largest Fe-Ti orebody in the RAP. In contrast to Tellnes, the intrusion contains magnetite as an important mineral phase. The deposit is a layered cumulate consisting essentially of plagioclase, ilmenite, orthopyroxene and magnetite (phimC). The intrusion has an arcuate shape and a length of approximately 4 km. Its thickness varies from 2-3 m at its eastern and western extremities up to a maximum of about 50-60 m in its central part. The strike of the modal layering is parallel to the country rock contacts and dips around 50° to the north in the central part of the intrusion. The dip gradually decreases with depth. Cumulate textures are overprinted by a tectonic fabric, which implies that the Storgangen intrusion had solidified prior to the doming of the Åna-Sira anorthosite. Layering is generally parallel with the contacts to the anorthosite and in the western extremity also with the layering in the Bjerkreim-Sokndal Layered Intrusion. It is therefore inferred that Storgangen was originally emplaced as a sub-horizontal sill-like body.

MgO content of ilmenite does, however, show a systematic variation with lower values generally found at increasingly marginal sites, and ilmenite and silicate compositions show an overall correlation with the modal amount of ilmenite (Robinson et al., this volume).

The ore deposit has been interpreted as having been emplaced as a noritic crystal

Mining has produced a 70 m-deep gorge that provides sections across the intrusion. The rock types range from anorthosite and leuconorite to norite and melanorite with oxide contents varying from near-zero to more than 50%. Massive ilmenite ore with anorthosite xenoliths is found at the footwall contact. The average TiO_2 content of the Storgangen cumulates is around 16%.

A slight cryptic layering defined by plagioclase and orthopyroxene compositions was noted by Krause et al. (1985) indicating more-differentiated cumulates towards the hanging-wall contact. However, the tendency for an up-section decrease in both anorthite and enstatite contents is subtle due to large compositional variability. Plagioclase and orthopyroxene compositions range from An_{52} and En_{72} to An_{46} and En_{66}, respectively. MgO contents of ilmenite closely mimic the compositional variation of orthopyroxene and vary from 2 to 3.5%. Chromium contents vary between 230 and 690 ppm Cr_2O_3 (Duchesne 1999, Duchesne & Schiellerup 2001). In addition, the amount of hematite exsolved from ilmenite is significantly reduced compared with the more primitive deposits.

The Storgangen intrusion, as a four-phase cumulate assemblage, cannot be an immediate differentiation product of the melt that formed the mono-mineralic anorthosite, unless the intermediate cumulates are hidden at depth. However, the presence of similar plagioclase megacrysts in both Storgangen and the Åna-Sira anorthosite, the occurrence of anorthositic layers in Storgangen, the contact relations and the joint post-magmatic deformation history suggest a close temporal and genetic link between the two units.

From 1916 to 1965, when production had irrevocably been transferred to the Tellnes deposit, Storgangen produced around 10 Mt of ore. However, with 60 Mt of documented reserves, Storgangen remains a deposit of world-class significance.

Other phimC deposits

Other intrusions in Rogaland that contain ilmenite-rich cumulates consisting of plagioclase, orthopyroxene, ilmenite and magnetite are present in the southern part of the Bøstølen Intrusion in the central part of the Åna-Sira anorthosite (Fig. 1,

#16) and parts of the Bjerkreim-Sokndal intrusion. Bøstølen is probably the oldest major intrusion in the Åna-Sira anorthosite, and is cross-cut by both norite pegmatites of the Blåfjell-Laksedal system and later mangerite dykes. The oxide-rich sections of the layered Bøstølen Intrusion have never been exploited, but average around 14% TiO_2 in the southern part of the intrusion. A small ilmenite deposit was exploited at Årstad in the southern part of the Bjerkreim-Sokndal Layered Intrusion (Fig. 1, #39) at the turn of the 19th century. This deposit contains plagioclase, orthopyroxene, ilmenite and magnetite, although it is located in a zone of phiC close to the base of the Layered Series.

In the Håland anorthosite along its contact with the Egersund-Ogna anorthosite, there are a string of deposits in a zone extending from the south end of Eigerøy eastwards for almost 10 km (Fig. 1, #50-80). These deposits are variably deformed but generally interpreted as cumulates (Duchesne 1999). In some cases, such as the deposits along the north bank of lake Kydlandsvatnet (#68-80), the orebodies may be parts of deformed layered norite sills, while mineralizations south of Koldal farm are more clearly related to norite dykes. More than 30 deposits have been exploited along this zone and the majority contain magnetite. Most of these deposits can be interpreted as phimCs. Some of the deposits north of lake Kydlandsvatnet do not contain magnetite and classify as phiCs while others (e.g. Rødemyr, Kaknuten) contain apatite as well magnetite. Duchesne (1999) reported ilmenite compositions in the phi±m deposits of this area with MgO ranging from 1.9 to 3.7% and Cr_2O_3 contents between 100 and 690 ppm. Further deposits with similar characteristics include some of the deposits at Laksedal in the Åna-Sira anorthosite (Fig. 1, #18) and Vatland in the Helleren anorthosite (Fig. 1,#45).

Apatite-bearing deposits

Apatite-bearing ilmenite deposits can be roughly divided into either nelsonitic occurrences (consisting principally of oxides and apatite) or layered cumulates of plagioclase, two pyroxenes, ilmenite, magnetite and apatite (phcimaC).

Allthough significant apatite is reported in a number of deposits (e.g. Rødemyr,

Kaknuten (Duchesne 1999)) and stray apatite grains may be found even in occurrences of pure ilmenite, nelsonite is relatively rare in Rogaland compared with other anorthosite provinces. Two deposits (Eigerøy and Hestnes) close to the contact between the Håland-Helleren and Egersund-Ogna anorthosites south of Egersund contain insignificant amounts of silicates and have been classified as nelsonites (Duchesne 1999). Duchesne (1999) reported MgO contents of ilmenite in these deposits of 1.1-1.4%.

Apatite-bearing oxide-rich rocks are otherwise confined to the evolved gabbro-noritic portions of the Bjerkreim-Sokndal Layered Intrusion. Other apatite-bearing rocks, such as the persistent jotunitic dykes transecting the province (Vander Auwera et al. 1998), various layered jotunitic bodies, as well as the uppermost part of the Bjerkreim-Sokndal Layered Intrusion, are generally relatively poor in ilmenite.

The Bjerkreim-Sokndal Layered Intrusion

The Bjerkreim-Sokndal Layered Intrusion has been studied almost continuously since the late 19th century (Michot 1960, 1965, Duchesne 1987, Wilson et al. 1996). The intrusion is exposed in three different lobes; the Bjerkreim, Sokndal and Mydland lobes. All lobes contain ilmenite prospects and exploitation has taken place in at least two localities in the Sokndal-lobe (Årstad and Hauge – Fig. 1, #39,43). The most extensive stratigraphy is present in the Bjerkreim-lobe where a 7 km-thick Layered Series is developed. The cumulate sequence may be divided into a number of mega-cyclic units (MCUs) each resulting from fractional crystallization after recharge of the magma chamber. Seven such MCUs have been defined (from base to top): 0, IA, IB, II, III and IV (Wilson et al. 1996). In addition, each MCU may be divided into a number of zones (a-f) based on certain index mineral phases (Fig. 2).

Only the uppermost MCU (MCU IV) displays the complete a-f zonal sequence, but apatite-bearing e-zones are found in MCUs IB, III and IV. In the Sokndal-lobe all cumulates belong to MCU IV, and in the Mydland lobe, which has been enveloped by a subsequent intrusion (Bolle et al. 1997), the cumulate sequence is at least equivalent to MCU IV. The mining opera-

tion at Hauge in the Sokndal-lobe took place in very oxide-rich cumulates of MCU IVe and resulted in a 25 m-deep shaft. Ore samples have been reported to contain up to 26% TiO_2 (Carlson 1945). Other unexploited occurrences in the Sokndal-lobe include Bakka, Årstadøy and Drageland (Fig. 1, #37,43,44), which are all located in the apatite-bearing MCU IVe zone, and contain up to 14% TiO_2 in selected samples (Carlson 1945).

In the Bjerkreim-lobe two large prospects have been identified as possible combined resources for low-MgO ilmenite, apatite and vanadian magnetite. One is located roughly between Åsen farm (Fig. 1, #82) and the village of Bjerkreim, following the strike of the layering for up to 3000 m. Stratigraphically, these cumulates belong to MCU IBe. At Åsen farm this zone averages 8% TiO2, 3.5% P2O5, and carries around 11% magnetite with 0.93% V_2O_3 in a 45 m section (Schiellerup et al. 2001, Meyer et al. 2002). The other prospect is located in the basal part of MCU IVe on the eastern flank of the lobe (Tekse-Bilstad – Fig. 1, #85), and constitutes a 30-90 m-thick section which may be followed along strike for several kilometres. Average compositions are on the order of 7-8% TiO_2 and 4% P_2O_5 (Schiellerup et al. 2001). The occurrence contain about 8% magnetite with 0.9% V_2O_3. Similar occurrences are also found on a smaller scale in the MCU IIIe zone. The apatite-bearing ilmenite deposits and prospects generally contain ilmenite with low contents of MgO and Cr_2O_3 and the ilmenite is poor in hematite. In all the Bjerkreim-lobe e-zone occurrences the MgO content of ilmenite averages between 1 and 1.8% (Meyer et al. 2002).

Genesis

The Rogaland ilmenite deposits have been ascribed to a variety of origins from metasomatic (Michot 1956, Hubaux 1960) to magmatic (Duchesne 1972, Roelandts & Duchesne 1979, Duchesne 1999). The magmatic hypotheses have included liquid immiscibility and the existence of Fe-Ti-rich melts (e.g. Force 1991). Currently, with the possible exception of the nelsonitic occurrences, all the ilmenite concentrations are considered to be cumulates (Duchesne 1996, 1999, Duchesne & Schiellerup 2001). As indicated above there is a systematic variation in

ilmenite composition with the mineralogical paragenesis of the individual ore bodies. Generally, the content of exsolved hematite, Cr_2O_3 and MgO in ilmenite decreases as the number of associated minerals increases and their compositions become more evolved (Duchesne & Schiellerup 2001). The composition of ilmenite from deposits region-wide correlate with the compositions of ilmenite in cumulates of similar mineralogical character in the Bjerkreim-Sokndal Layered Intrusion. These observations support the suggestion of cumulus processes to explain the formation of most ilmenite deposits in Rogaland as well as a derivation from similar types of melt. Except for the Bjerkreim-Sokndal Layered Intrusion, which formed from melts differentiating in situ, parental melts must have evolved through fractional crystallization at depth. Discrepancies do exist among certain trace elements, most notably chromium, which may to some extent reflect variable pressure conditions during the genesis of the province and its ilmenite resources (Vander Auwera et al. 2000, Duchesne & Schiellerup 2001). The correlation of MgO in ilmenite with the magmatic evolution is important in prospecting for specific qualities of ilmenite, though other processes may also affect the compositions. Mg is the principal pollutant in ilmenite and the request for better quality ilmenite has led to the identification of lower-grade deposits containing low-MgO ilmenite with apatite and vanadian magnetite as additional resources (e.g. the Bjerkreim-Sokndal Layered Intrusion).

Whereas the mineralogical composition of the Rogaland ilmenite deposits is evidently controlled by fractional crystallization, the ilmenite-rich nature of the deposits is more difficult to explain. The composition of the Rogaland ilmenite deposits sensu strictu with TiO_2 contents in excess of 15%, such as Tellnes, Storgangen, Blåfjell, Vardåsen and other primitive deposits, seems to be strongly offset from the cotectic modal proportions of the crystallizing phases. The cotectic proportion of ilmenite (or TiO_2) in plagioclase-orthopyroxene-ilmenite cumulates may be estimated from the composition of the c-zones in the Bjerkreim-Sokndal Layered Intrusion. The c-zone cumulates are typically ad-cumulates and make up a considerable part of the Layered Series (Fig. 2). Eigh-

teen whole-rock analyses from the c-zones of MCU II and III yield an average of 4.4% TiO_2. This TiO_2 content is only marginally higher than in the inferred parental melt (3.46%) (Duchesne & Hertogen 1988), and is in agreement with the relatively small quantity of ilmenite produced in experimental runs on the parental composition (Vander Auwera & Longhi 1994).

Cumulates which are offset from cotectic proportions have been explained by magma mixing, either through the forcing of magma compositions away from cotectic equilibrium (e.g. Meyer & Wilson 1999) or by resorption-crystallization processes (Tegner & Robins 1996, Lundgaard et al. 2002). Similarly, the model for the formation of the Tellnes ilmenite deposit advanced by Robinson et al. (this volume) involves mixing of variably evolved magmas. In cases where extensive layering is present in the intrusions, such as in Storgangen, wholesale mixing processes are more difficult to envisage and probably require mixing in an open system with high magma throughput. Additional work should be undertaken in order to verify the nature of the processes that resulted in the pronounced ilmenite enrichment observed.

Sulphides

Although Fe-Ti ore is the foremost economic resource of the RAP, base metal sulphide deposits are found locally. The Fe-Cu-Ni sulphide occurrences may be divided into two types. **1:** minor anorthosite-hosted, disseminated to semi-massive deposits, and **2:** disseminated occurrences hosted by noritic or orthopyroxenitic cumulates.

In Rogaland, Fe-Cu-Ni sulphide deposits hosted by massif-type anorthosites are generally small and disseminated, although massive ores are also found. The largest known deposit is the Homsevatnet deposit (Fig. 1, #98) estimated at 75,000-100,000 tons of sulphide ore (Hovland 1975). Several other deposits were exploited on a small scale at the end of the 19th century, and during the 1970s and late 1990s there was renewed interest in the area and airborne geophysical prospecting was carried out. The anorthosite-hosted mineralizations are widely scattered in three of the major anorthosite bodies; the
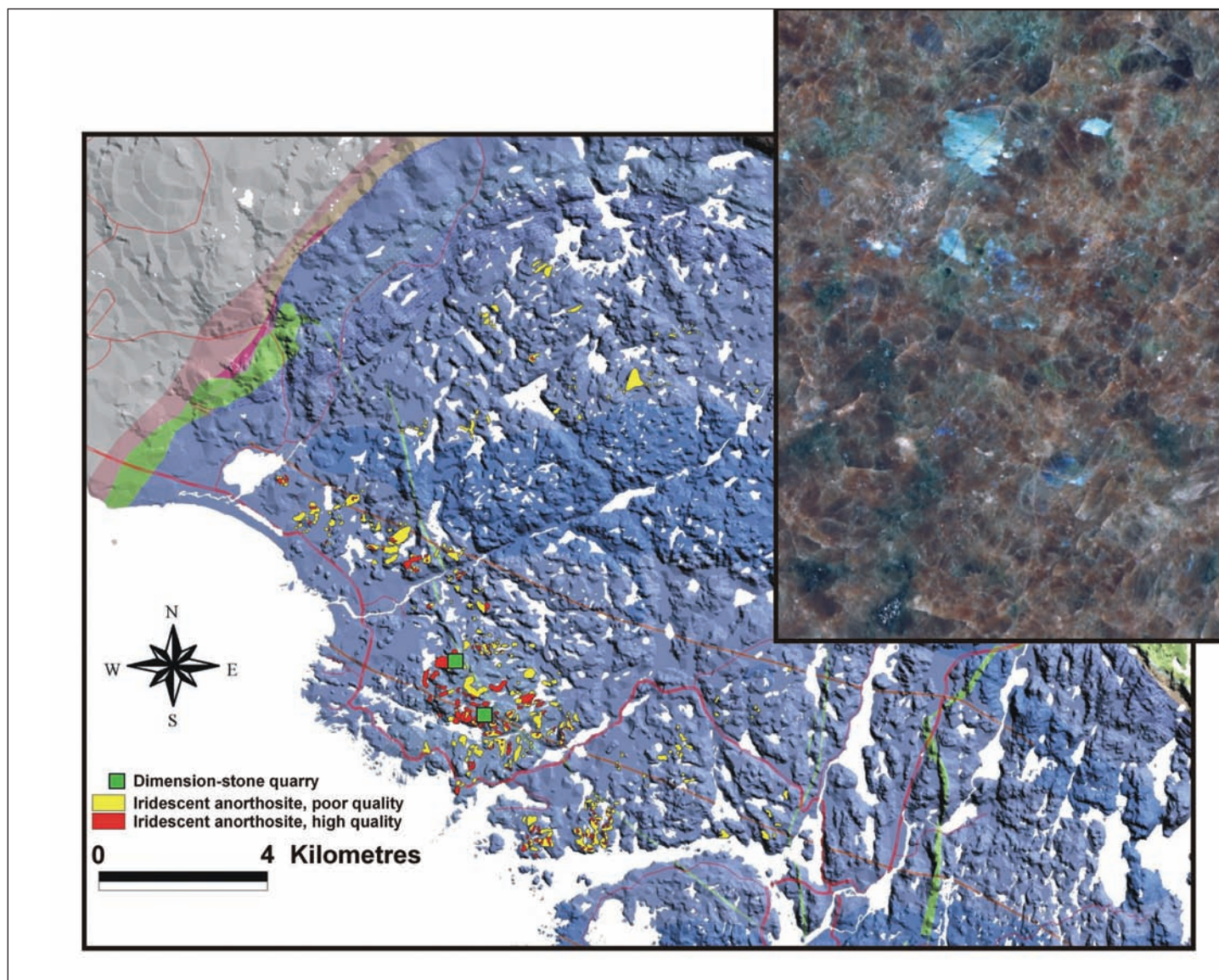
Fig. 3. Dimension stone quarries and occurrences of anorthosite with labrado-rescent plagioclase in the western part of the RAP. Anorthositic rocks are shown in blue, jotunitic rocks in green, basaltic dykes in red and the surrounding gneisses in pink and brown (see Marker et al. (2003) for details). Above right is shown a polished surface of typical coarse-grained anorthosite with iridescent plagioclase.

Egersund-Ogna, Håland and Helleren anorthosites (Fig. 1).

Although there is a variation in chemical composition, morphology and to some extent mineralogy among the deposits, they are remarkably similar in a number of ways. All investigated anorthosite-hosted sulphide occurrences are found in close spatial association with noritic or pyroxenitic dykes, pockets or lenses within the anorthosites. They are dominated by an assemblage of pyrrhotite, chalcopyrite, pentlandite and pyrite in accordance with a magmatic origin. No hydrous silicate minerals are associated with any of the deposits in the province.

A compilation of the known sulphide occurrences in the Rogaland province is presented in Table 2 and their locations are shown in Fig. 1.

Dispersed sulphides are ubiquitous in virtually all mafic (jotunitic-noritic) intrusions associated with the Rogaland anorthosites. Oxide-rich layered intrusions such as the Storgangen and Bøstølen intrusions as well as the Tellnes dyke generally contain 0.5-2 vol% sulphides, and Cu and Ni sulphides are recovered as a biproduct in ilmenite beneficiation. In the Bjerkreim-Sokndal Layered Intrusion sulphides are found as dispersed accessory constituents throughout the stratigraphy, but they also appear in a thin stratiform enrichment, more than 30 km in lateral extent (Fig. 1, #100). The sulphides are hosted by an up to 3 m-thick layer of orthopyroxenite or melanorite located at the boundary between MCU II and III (Fig. 2). The sulphides are interstitial and the sulphide content is up to about 9%. Its formation has been attributed to a mixing event between a bouyant primitive mag-

ma and a more evolved resident magma, both saturated in sulphide. The resulting hybrid magma crystallized an assemblage dominated by orthopyroxene and ilmenite which sank to the magma chamber floor as dense plumes containing exsolved sulphide melt (Jensen et al. 2003).

The sulphide paragenesis is similar to that observed in the anorthosite-hosted deposits. The sulphide assemblage is dominated by pyrrhotite and euhedral pyrite grains are often embedded in chalcopyrite. As at Homsevatnet, violaritization of precursor pentlandite is extensive and in several samples pentlandite has been completely replaced. 'Birds eye' replacements of pyrrhotite by marcasite-pyrite aggregates are common.

Sulphides contained in other mafic, dominantly noritic intrusions are different only in modal amounts and the extent of replacement and alteration. The sulphide assemblage is always found to be pyrrhotite, pentlandite, chalcopyrite and commonly, but not invariably, pyrite. In several intrusions, including the Bjerkreim-Sokndal Layered Intrusion, violaritization may be accompanied by a more severe alteration of pyrrhotite to fine-grained aggregates of euhedral pyrite and magnetite. In intrusions such as Tellnes, Storgangen or Bøstølen, this reaction has created a sulphide paragenesis commonly dominated by fine-grained pyrite and locally almost devoid of pyrrhotite. Because the pyrite-magnetite aggregates have never been observed in pentlandite-bearing samples and only in violaritized samples, it is inferred that the breakdown of pyrrhotite succeeds the breakdown of pentlandite.

In parts of Tellnes and locally in Storgangen, millerite is the nickel-bearing sulphide phase. In these cases millerite is invariably closely associated with grains of siegenite as the cobaltian phase. Because millerite-bearing samples are devoid of pyrrhotite (this having been replaced by pyrite and magnetite), the millerite-siegenite assemblage is considered part of a late-stage, low-temperature (non-supergene) paragenesis. The occurrence of millerite as the nickel-bearing sulphide phase in Tellnes results in a high Ni content in the recovered sulphide concentrate.

Recalculated to 100% sulphides, the Rogaland occurrences contain up to 2.7% Cu and 4.2% Ni and no more than 0.5% Co. Noble metal concentrations are very low, generally below 10 ppb. Pt concentrations are lower than Pd, and common Os tenors (Os concentration in the sulphides at the time of formation) range from 0.06 to 14.7 ppb (Schiellerup et al. 2000).

Genesis

The sulphide assemblage and the absence of hydrous mineral phases associated with the sulphides strongly suggest a magmatic origin for all the sulphide occurrences. Schiellerup et al. (2000) argued that the Rogaland sulphide occurrences are isotopically indistinguishable from both the hosting anorthosites and the associated mafic intrusions, and that a common source had to be invoked. As discussed by Schiellerup et al. (this volume), the noble and chalcophile element variation also suggests a common heritage of sulphides hosted by anorthosites and mafic intrusions, probably involving similar parental magmas. In their model all chalcophile element variation may be explained by different R-factors (R-factor: ratio between amounts of silicate melt in effective equilibrium with the sulphide melt) and a fairly well-defined parental sulphide composition. In particular, the low noble element contents suggest a derivation from noble element depleted magmas, and that low R-factor processes were at play in the generation of most anorthosite-hosted deposits. This implies a more or less static exsolution of a sulphide melt, and that the exsolution most likely resulted from the differentiation of the parental magma. In the mafic intrusions the fairly constant sulphide content in the cumulates indicates that continuous segregation from sulphide-saturated melts took place and R-factors are therefore only slightly elevated. The main exception in terms of grade is the Bjerkreim-Sokndal Layered Intrusion, where the stratiform sulphide enrichment is related to a magma replenishment event involving magma mixing (Jensen et al. 2003).

Dimension stone

The first exploitation of anorthosite for dimension stone within the RAP took place at the beginning of the 20th century. At that time quarrying was mainly aimed
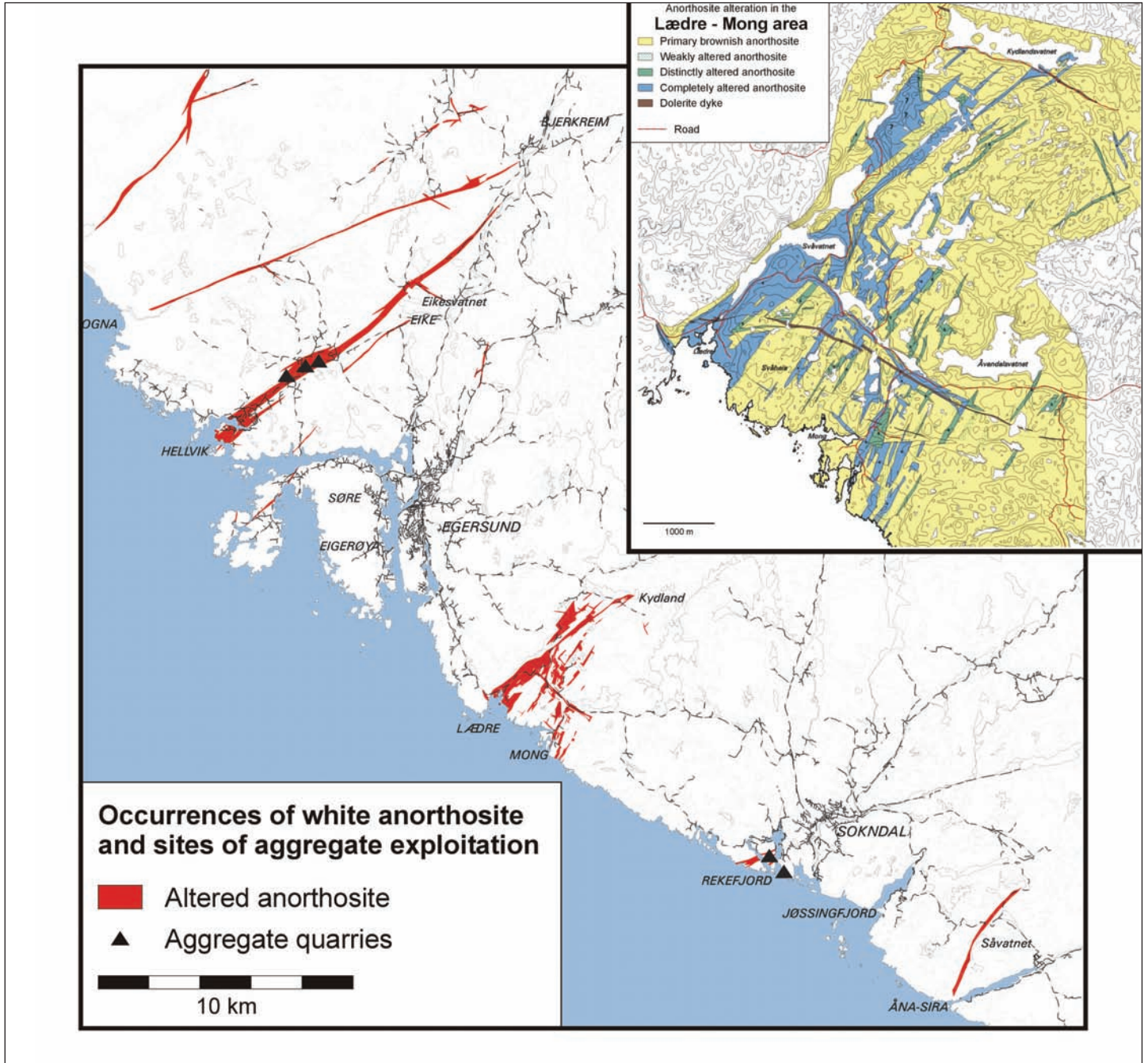
Fig. 4. Map of the occurrence of white altered anorthosite in the RAP, with the locations of current hard-rock aggregate production indicated. Inset map shows the Lædre-Mong alteration zone in detail.

at local housing. Industrial-scale production did not start until the mid 1990s, based on regional and detailed geological mapping carried out by the Geological Survey of Norway in co-operation with the local governments of the region (Heldal & Lund 1994). Dimension-stone quality anorthosite contains iridescent plagioclase (Bøggild intergrowths), displaying a green, yellow or blue play of colour (Fig. 3). This type of anorthosite is highly attractive on the international market, and can be compared with the better known larvikite in the Permian Oslo-rift.

The attractive anorthosites occur predo-

minantly in the central part of the Egersund-Ogna Massif, in the Hellvik area. In addition to being enriched in iridescent feldspar, these anorthosites contain megacrysts of orthopyroxene. The deposits occur as scattered, irregular blocks, surrounded by megacryst-poor anorthosite containing a less calcic plagioclase. The clustering of the more mafic fragments in the central and thus lower part of the dome-shaped Egersund-Ogna Massif may indicate that their origin relates to the earliest stage of crystallization of the massif. At a later stage of the magmatic evolution, these early-crystallized rocks have been disrupted and emplaced during

the diapiric uprise of the partially consolidated magma.

The highly irregular nature of the occurrences makes volumetric interpretation and reserve estimates difficult without coring and trial extraction. In addition to the size of the deposits, the market value, and thus the exploitability, depends on several factors; it must be possible to produce large blocks (lack of penetrating fracture systems), hydrothermal alteration that causes discolouring of the rock should not be present, the colour and structure should be homogeneous (predictable high quality) and finally the geometry of the deposits must be suitable for an efficient quarry layout.

In recent years, detailed mapping has been carried out of the valuable anorthosite types and has defined a large number of occurrences within five to six principal areas (Fig. 3). Two companies are now producing high-value blocks in the area, and it is expected that production will steadily increase in the future. The current export of anorthosite blocks from the two quarries amounts to around 6.5 million _ per year. At the present time, an effort is being made to secure the anorthosite resources for the future by integrating the resource maps into land use planning.

Hard-rock aggregate

Aggregate exploitation in Rogaland is taking place in zones of altered anorthosite, leuconorite or jotunite. The Rogaland province contains several zones in which the primary purple-brown anorthosite has been partly or completely altered by hydrothermal solutions to white anorthosite, which is attractive for the hard-rock aggregate industry. Alteration tends to follow certain prominent topographic lineaments and affects all rock types crossed by the lineaments. However, it is primarily the altered anorthosite and leuconorite which is attractive because of the improved mechanical properties (in particular the wearing and polishing properties) combined with pale greyish to white colours and lack of quartz. The degree of alteration varies within the zones, but even partial alteration improves the mechanical properties considerably. Most attractive is completely-altered white anorthosite without mafic minerals,

which also has a potential for more valuable industrial products such as fillers. An additional advantage for the aggregate industry is the localization in the coastal zone with good discharging possibilities.

Anorthosite and leuconorite occupy large areas of the western and southern parts of the RAP. Recent mapping has shown that within this area, altered white anorthosite occurs in six principal zones that are separated by large areas of primary anorthosite with insignificant alteration (Fig. 4). All of the main zones are steeply dipping and have a NE-SW orientation following marked valleys in the terrain. Associated with the main zones there are subordinate alteration zones, some with a different orientation making the internal structure of the zones very complex. There is a tendency that the extent and intensity of the alteration are largest where subordinate zones join or cross a main zone. From southeast to northwest, the six main zones are: The Åna-Sira, Rekefjord, Lædre, Hellvik, Ogna and Brusand zones (Fig. 4). Extraction currently takes place in the Rekefjord and Hellvik zones with a total annual extraction of 1.3 million tons (2001) of which most is exported. The production value amounts to around 8 million € per year.

The white anorthosite in the alteration zones was formed by hydrothermal activity at a time when the rocks in the RAP were solidified and the structure established. This is evident because the original textures and structures are preserved in the hydrothermally altered anorthosites. No destruction of the original textures and structures is observed in the steep, cross-cutting alteration zones showing that alteration was not accompanied by displacement.

As shown in Fig. 4, the principal alteration zones generally trend NE-SW. Another important trend is NW-SE, as in the Lædre zone, while N-S trends are less prominent. These trends coincide with the main joint directions that are observed in the province. Because of the lack of evidence for movement, it is likely that the hydrothermal alteration propagated along zones of weakness in the solidified anorthosite body and was later overprinted by jointing in the same directions. The alteration and introduction of hydrothermal

Deposit	Major phases						Significance
	ilm	plag	opx	mt	ap	cpx	
Pramsknuten/Jerneld	x						Minor
Svånes	x	x					Minor
St. Ålgård	x	x					Minor
Florklev	x	x	(x)				Minor
Blåfjell	x	(x)	(x)				Significant
Vardåsen	x	x	x				Minor
Raunsli	x	x	x				Minor
Frøytlog	x	x	x				Minor
Beinskinnvatnet	x	x	x				Minor
Årstad	x	x		x			Minor
Storgangen	x	x	x	x			Major
Bøstølen	x	x	x	x			Minor
Tellnes	x	x	x	(x)			Major
Laksedal	x	x	x	(x)			Significant
Koldal field	x	x	x	(x)			Minor
Vatland	x	x	x	(x)			Minor
Fisketjern	x	x	x	x			Minor
Kydlandsvatnet field	x	x	x	(x)	(x)		Minor
Eigerøy	x			x	x		Minor
Hestnes	x			x	x		Minor
Kaknuden	x	x	x	x	x		Minor
Rødemyr	x	x	x	x	x		Minor
Drageland	x	x	x	x	x	x	Minor
Hauge	x	x	x	x	x	x	Significant
Bakka	x	x	x	x	x	x	Significant
Åsen	x	x	x	x	x	x	Significant
Tekse-Bilstad	x	x	x	x	x	x	Significant

Table 1. Key ilmenite deposits in the RAP according to phases present and significance. Partly based on Duchesne (1999).

fluids must be older than the basaltic dykes of the 616 Ma old Egersund dyke swarm (Bingen at al. 1998) that cut the altered anorthosites in the Lædre area but are themselves unaffected.

During the hydrothermal alteration the plagioclase is albitized with formation of epidote, clinozoisite and small amounts of white mica (sericite). These secondary minerals form a fine-grained aggregate within the outlines of the original plagioclases. The primary orthopyroxene in the unaltered anorthosite is transformed into amphibole and eventually decomposed into chlorite and epidote/clinozoisite in the highly altered anorthosite. At the same time ilmenite is progressively transformed into rutile, or further into titanite.

The altered white anorthosites show markedly improved mechanical properties compared to the primary purple-brown anorthosite. The standard test methods show that the white anorthosites have properties that fall into the best class of hardrock aggregates (Erichsen & Marker 2000). White anorthosites are being quarried for aggregate in the Hellvik and Rekefjord zones, but the aggregate potential is much larger in the area. This concerns not least the, as yet, unexploited Lædre zone (Fig. 4), where reserves of white anorthosite are among the largest in the Rogaland Anorthosite Province.

Kaolin

An occurrence of kaolin was discovered on the farmland at Dydland, southeast of Sokndal, around 1880, and in 1898 a new company, Den Norske Chamottefabrik, acquired the rights to exploit the occurrence. The kaolin is located in a narrow east-west striking gully and, according to Reusch (1900), can be followed for a length of about 1 km with a width of approximately 10-20 m. JHL Vogt (in Carlson 1945) and H. Reusch (1900) associated the kaolinitization with one of the basaltic Egersund dykes striking parallel with the occurrence, and interpreted the alteration as a result of hydrothermal activity along the contacts between dyke and anorthosite. The spatial association with the ESE-WNW striking basalt dykes implies that the kaolinitization is not related to the alteration that resulted in 'white anorthosite'.

The main site of exploitation was about 20 m long, 5 m wide and up to 3 m deep. Today the deposit is exhausted and the kaolinitization mainly evident through the occurrence of a light coloured soil.

Conclusions

The Rogaland Anorthosite Province in Southwest Norway contains some of the largest mineral resources in Norway with on-going mining of ilmenite, dimension stone and aggregate. In all, the three types of geo-resources represent a yearly revenue exceeding 75 million €. Active exploration has located several additional prospects of aggregate and dimension stone, and alternative titanium resources may, in the future, be mined in conjunction with apatite and vanadian magnetite. A titanium resource in the Storgangen intrusion has been abandoned for several deca-

Table 2. Compilation of known anorthosite-hosted sulphide deposits in the RAP. 1) Stadheim (1938). 2) Henriette (1986). 3) Geological Survey of Norway, 'Bergarkivet'. 4) Jensen et al. (2003). The Bjerkreim-lobe is the northern lobe of the Bjerkreim-Sokndal Layered Intrusion. See Fig. 1 for locations.

Host unit	Occurrence	Characteristics	Host rock
Egersund-Ogna	Homsevatnet	Massive dyke-like or lenticular body dominated by po.	Norite pegmatite.
	Høgatuva[1] (Ualand)	Coarse-grained sulphides in irregular lenses or laminae up to 3 m long.	Not known.
	Fossåsen[1] (Ualand)	Sulphide impregnation.	Not known.
	Tunggårdsbakken[1] (Ualand)	Irregular lens or dyke up to 1.60 m wide containing coarse-grained sulphides.	Not known.
	Bjørndalsnipa	Three cm/dm-sized zones of semi-massive po-dominated ore, surrounded by cp-dominated veins or disseminations.	Mineralizations hosted by anorthosite, but pegmatoid norite is present in cored secton.
	Myklebust	Disseminated sulphides and irregular veins dominated by po.	Host rock mostly noritic or pyroxenitic.
	Fossfjellet	Dyke or vein type po-dominated ore surrounded by cp-dominated disseminated sulphides.	Strongly foliated norites and anorthosites in the deformed margin of the anorthosite.
	Piggstein[2]	Disseminated po-dominated sulphides.	Mafic pyroxenitic (opx+cpx) part of noritic lens or dyke.
	Helland[3]	Contains Ni-pyrrhotite. Details not known.	Not known.
	Hellvik[3]	Contains po-pn, cp, py. Details not known.	Not known.
	Skora[3]	Contains Ni-pyrrhotite. Details not known.	Not known.
Håland-Helleren	Urdal	Dyke-like sulphide body consisting of a central massive po-dominated zone surrounded by veinlets of sulphides richer in cp.	Main mineralization hosted by anorthosite, but located within metres of a norite dyke.
	Koltjørn[3]	Contains pyrrhotite (Ni). Details not known.	Not known.
	Grastveit[3]	Contains Ni-pyrrhotite. Details not known.	Not known.
Bjerkreim-lobe	Rudleknuten[4]	Stratiform disseminated sulphide concentration in layered cumulates of the Bjerkreim-Sokndal intrusion.	Orthopyroxenite or melanorite.

des but still contains large reserves. Historically, small-scale mining for Ni and Cu sulphides as well as for kaolin, has been of local importance.

Rogaland, as a classic massif-type anorthosite province, thus provides the complete range of resource types usually associated with this type of setting. The ilmenite resources are primarily controlled by cumulate processes and emplacement mechanisms, and to a large extent have predictable compositions. The occurrence of high-quality dimension stone with iridescent plagioclase is related to the composition of the plagioclase and to the structural habit of the host anorthosite. The location of desirable aggregate prospects and extraction sites, as well as of kaolin, are controlled by zones of alteration along major fracture zones or dykes. Today, the current extraction, exploration and future prospects rank the Rogaland Anorthosite Province as one of the foremost mineral resource provinces in Norway.

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Titanium and Iron-Titanium Deposits in Norway

Approximately 300 titanium and iron-titanium ore deposits are known in Norway; most of these occur in distinct Proterozoic and Phanerozoic geological provinces.
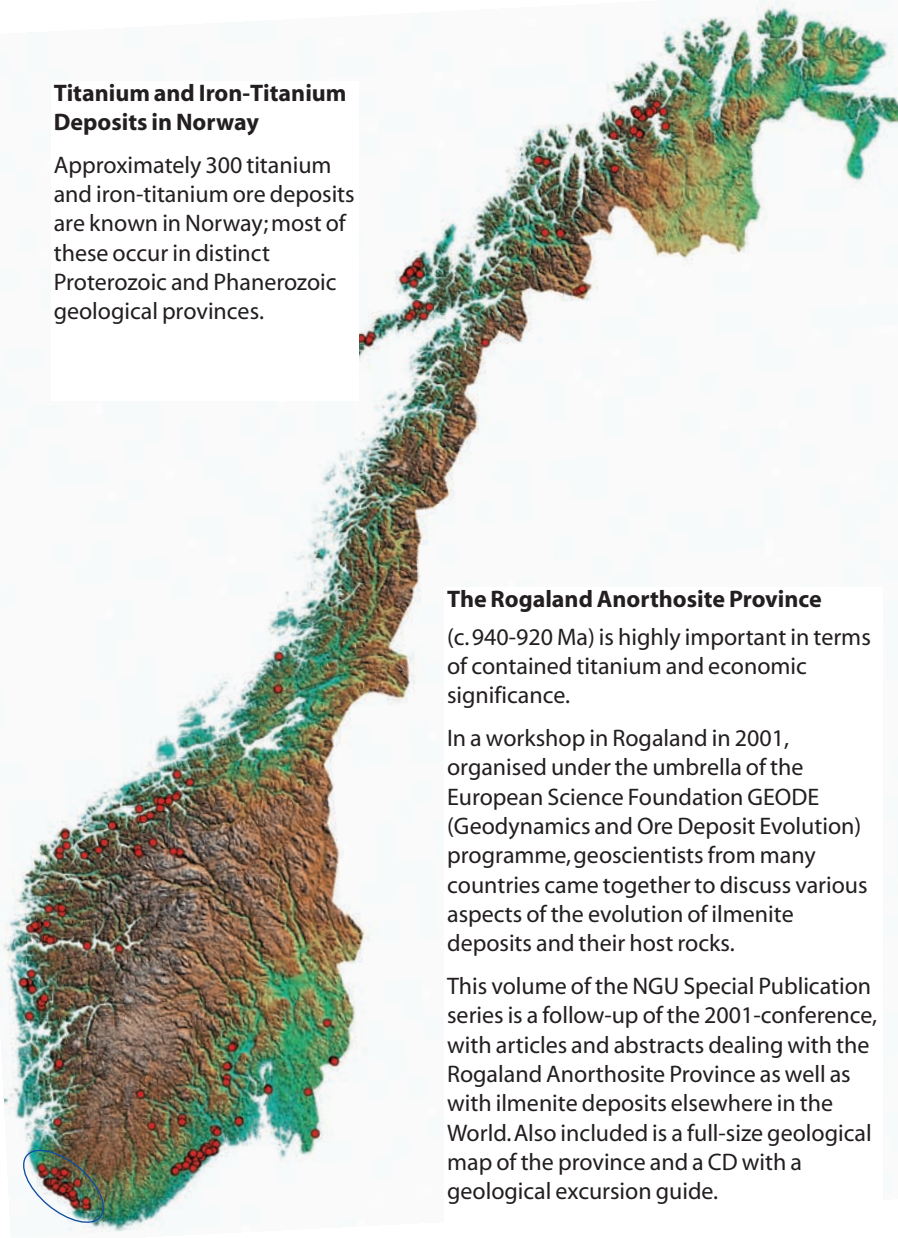
The Rogaland Anorthosite Province

(c. 940-920 Ma) is highly important in terms of contained titanium and economic significance.

In a workshop in Rogaland in 2001, organised under the umbrella of the European Science Foundation GEODE (Geodynamics and Ore Deposit Evolution) programme, geoscientists from many countries came together to discuss various aspects of the evolution of ilmenite deposits and their host rocks.

This volume of the NGU Special Publication series is a follow-up of the 2001-conference, with articles and abstracts dealing with the Rogaland Anorthosite Province as well as with ilmenite deposits elsewhere in the World. Also included is a full-size geological map of the province and a CD with a geological excursion guide.

Rogaland
Anorthosite Province

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