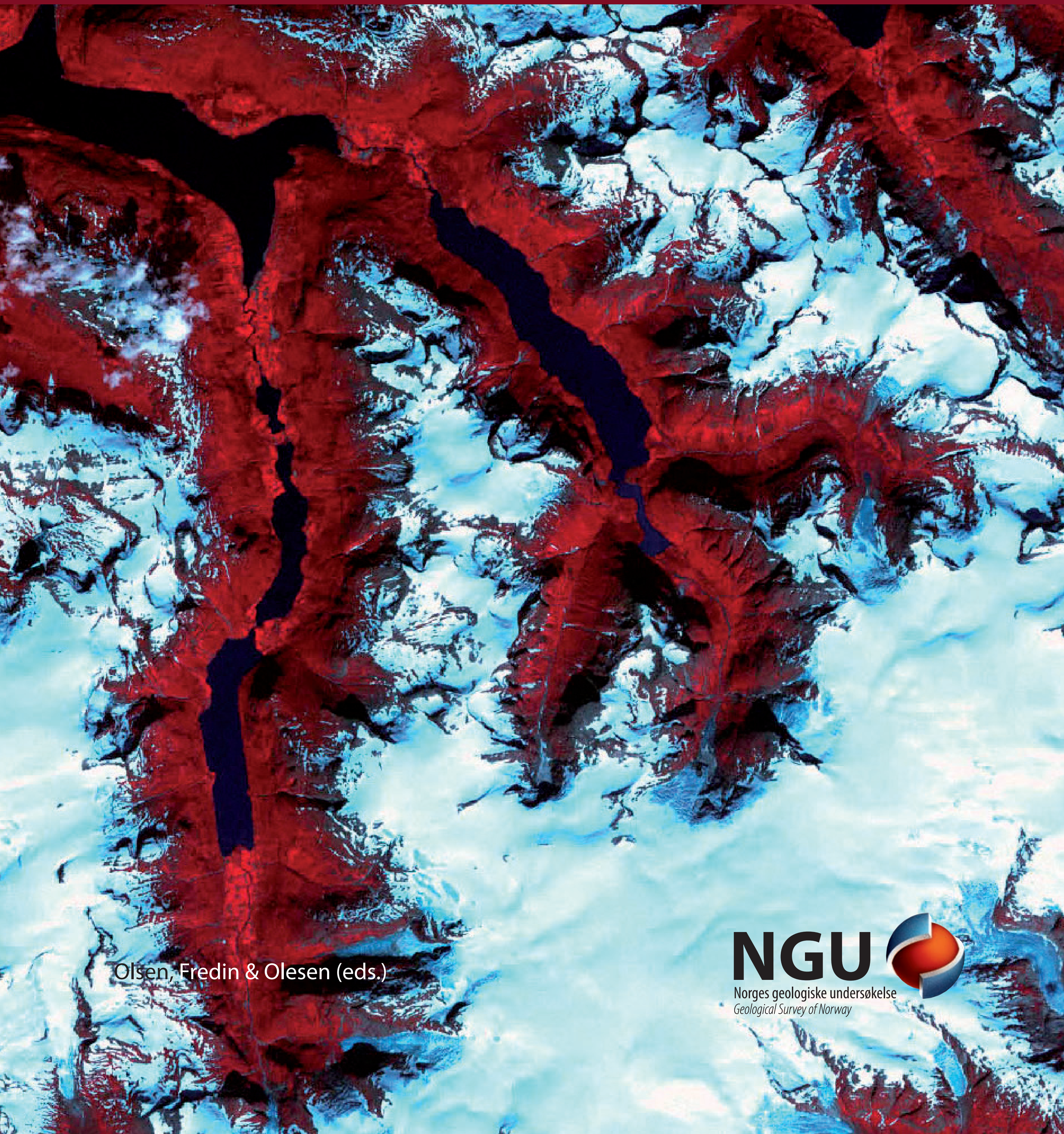
QUATERNARY GEOLOGY OF NORWAY

Geological Survey of Norway Special Publication, 13



Olsen, Fredin & Olesen (eds.)

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Introduction

The profound impact of glaciers on the development of the landscape in Norway has become common knowledge among most geologists during the last centuries. For example, the alpine landscape, the U-formed valleys with deep basins and bedrock thresholds, the fjords, the strandflat, the glacially polished and striated rock surfaces, and the various and numerous glacial deposits all bear clear evidence of strong glacier influence. Glaciers have advanced and retreated in Norway and adjacent areas a number of times during the last 2–3 million years, and therefore it is commonly thought that the present-day landscape was formed mainly during this very young period of the geological time scale.

There are, however, several preglacial landscape features that are preserved or only slightly changed during these young glaciations. Both types of landscape, i.e., the glacially formed landscape and the preglacial landscape, are presented in the first chapter of this volume. This chapter deals with both the processes and their products in the history of landscape development. Starting with the preglacial development, it continues through the ice ages, and ends with processes and landscape development in the Holocene time period.

The second chapter discusses the glacial history of Norway during the Quaternary time period, and describes the glacier variations with advance and retreat of the glaciers and ice sheets.

The third chapter presents the record of deposits from the represented stadials, interstadials and interglacials during the last 300,000 years in Norway and adjacent areas, and the suggested correlations between these deposits.

The fourth chapter deals with neotectonics in Norway and adjacent areas, which are also largely a result of ice-sheet advance, erosion, sedimentation and retreat. The pressure variations on the ground surface as a result of variable ice weight, glacier erosion and accumulation have led to many earthquakes and large faults, such as the postglacial Stuoragurra fault on Finnmarksvidda.

This compilation of the present knowledge of the Quaternary geology of Norway will be of use to several of our fellow colleagues of geology, students, and other dedicated, non-professional readers.

Trondheim, December 2012

Lars Olsen   Ola Fredin   Odleiv Olesen

Glacial landforms and Quaternary landscape development in Norway

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The Norwegian landscape is a function of geological processes working over very long time spans, and first order structures might for example be traced to ancient denudational processes, the Caledonian orogeny or break-up of the North Atlantic. However, a large portion of large-, intermediate-, and small-scale landforms in Norway owe their existence to Quaternary glaciations or periglacial processes. The Quaternary time period (the last *c.* 3 million years), is characterised by cool and variable climate with temperatures oscillating between relative mildness to frigid ice-age conditions. A wide spectrum of climate-driven geomorphological processes has thus been acting in Norway, but the numerous glaciations have had the most profound effect with the production of large U-shaped valleys, fjords and Alpine relief. On the other hand, interior and upland areas in Norway seem to be largely unaffected by glacial erosion and exhibit a possibly pre-Quaternary landscape with only some periglacial influence. The ice sheets in Scandinavia thus have redistributed rock mass and sediments in the landscape with the largest glaciogenic deposits being found on the continental shelf. Large Quaternary deposits and valley fills can also be found onshore and these are now valuable resources for aggregates, ground water and agriculture. Important Quaternary processes have also been acting along the Norwegian coast with denudation of the famous strandflat, where the formation processes are not fully understood. The isostatic depression of crust under the vast ice sheets have also lead to important consequences, with thick deposits of potentially unstable, fine-grained glaciomarine sediments in quite large areas of Norway. It is thus clear that much of Norway's beauty but also geohazard problems can be explained with the actions of Quaternary geomorphological processes.

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Introduction

The dramatic landscape of Norway is to a large extent formed by surface processes during the Quaternary time period, most notably by the action of glaciers through numerous glaciations. As glaciers move across the Earth's surface they erode and transport sediments. Ice sheets and glaciers are thus regarded as powerful agents of erosion, which is manifested in the grandeur of e.g., glacial fjords and valleys. A Norwegian fjord can thus bear witness to glacial erosion on the order of 2 km (Nesje and Whillans 1994, Kessler et al. 2008). However, in the last decade it has also been generally accepted that glaciers and ice sheets may rest on the ground for long periods of time without affecting the landscape. In fact, a nonerosive ice sheet may even protect delicate landforms from subaerial processes. This perception has opened up new perspectives for glacial geologists and gemorphologists who seek to understand landscape development, since landforms and sediments may survive glacial overriding. Instrumental for understanding the Quaternary landscape development has also been the emerging body of global ice-volume proxies (e.g., Martinson et al. 1987), which have allowed glacial geologists to explore the different modes of glaciations that have affected the landscape (Porter 1989, Stroeven and Kleman 1997, Fredin 2002). Indeed it has been shown that up to 40 glacial cycles have waxed and waned in Scandinavia during the last million years. During a large portion of the Quaternary, glaciers and ice caps have been restricted to the mountain areas and occasionally migrated down to the lowlands and coasts of Scandinavia (Porter 1989, Fredin 2002). Only during the last ~700 kyr, full-scale glaciation cycles have been operating such that the dominant mode of glaciation is centred around the mountain chain, where also the main erosion has taken place (Storeven and Kleman 1997, Figure 1). During numerous glaciations, ice divides, ice-marginal-, erosional-, and depositional zones have been migrating and affected the landscape at different locations and at different time periods (Kleman et al. 2008). Old landforms and deposits may thus have been preserved, obliterated or reworked. At a regional scale, the Quaternary landscape that we see today is the cumulative result of processes acting during numerous glacial and interglacial periods. On the other hand, local-scale features, in particular depositional glacial landforms, can often be attributed to the Weichselian glacial cycle and the last deglaciation.

A typical Norwegian alpine landscape, e.g., in Jotunheimen or the Lyngen Alps, thus is a mosaic of landforms with widely different ages closely juxtaposed to each other. The youngest landforms postdate the last main deglaciation and typically consist of late glacial or Holocene landforms such as moraines and meltwater channels, colluvial landforms such as talus and debris cones, fluvial landforms such as deltas or alluvial plains, and coastal landforms such as wave cut benches and raised beaches. Underneath the postglacial landforms and sediments, or close by, we find major glacial landforms such as drumlins, eskers and end moraines. All these landforms may be sitting in a glacial fjord that has been eroded through the cumulative action of many glaciations.

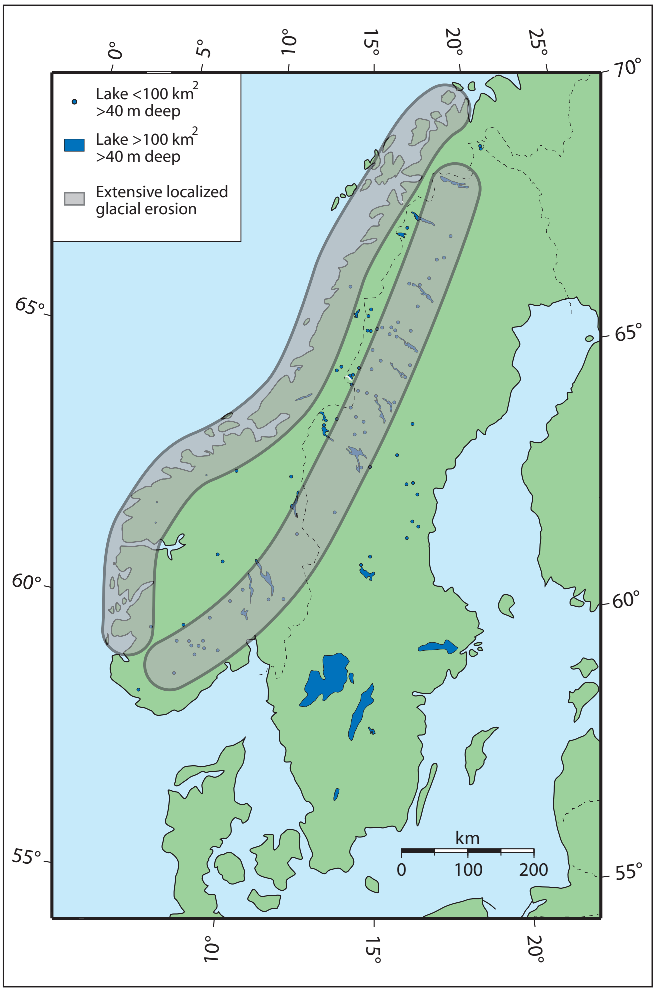


Figure 1. First-order glacial erosion landforms are centred around the Scandinavian mountain chain with fjords on the Norwegian coast and deep piedmont lakes east of the mountains

Major advances in glacial geomorphology during the last two decades are founded on a greater understanding of glacier mechanics and glacier thermal distribution (cf., Paterson 1994). The ability of glaciers to construct, erode, reshape and preserve landforms is to a large extent dependent on the basal temperature of the ice mass (Kleman and Hättestrand 1999), such that glacial erosion scales to subglacial sliding velocity (Kessler et al. 2008). A basal ice temperature at the pressure melting point allows the glacier to slide on the substratum and more importantly permits subglacial sediments to deform (Hooke 1989, Paterson 1994, Hooke 1998). By contrast, a glacier where the basal layers are below the pressure melting point can only slide very slowly or not at all on the substratum and the underlying sediments, which also are frozen, and are more competent than the glacier ice. Hence, these sediments cannot be deformed or eroded by the glacier ice. An important consequence of this is that landforms and sediments (glacial and nonglacial) may survive ice-overriding, a notion not widely accepted only two decades ago (Sollid and Sørbel 1994). Furthermore, this understanding has allowed research to not only focus on the last glacial maximum and deglaciation but on the whole last glacial cycle and beyond (Kleman et al. 1997, Boulton et al. 2001).

In addition to better understanding of glacial processes and landform genesis, new advances in dating techniques, most notably in situ terrestrial cosmogenic dating and luminescence dating have allowed researchers to pinpoint the age of previously elusive Quaternary landforms and sediments (e.g., Gosse and Phillips 2001, Duller 2004). This has greatly aided Quaternary geologists to put previously undated landforms into a time frame and has helped reconstruct the Quaternary geology of Norway and other formerly glaciated areas.

Pre-Quaternary landforms

Upland areas in Norway and Scandinavia are often surprisingly flat and show few signs of glacial alteration, which is in stark contrast to neighbouring deeply cut glacial valleys (Figure 2). This conception was early developed by e.g., Reusch (1901), Ahlmann (1919), Holtedahl (1960) and Gjessing (1967). Based on geomorphological evidence these authors generally argued for a late uplift, Neogene, of the Scandinavian mountain range. The detailed mode and configuration of this uplift is the topic of another chapter in this volume (Olesen et al. this issue). There has been a recent surge in interest for remnants of pre-Quaternary landscapes since they convey information about geological and climatological conditions predating the ice ages and also Quaternary ice dynamics (Nesje and Dahl 1990, Kleman and Stroeven 1997, Lidmar-Bergström et al. 2000, Bonow et al. 2003, Fjellanger 2006, Fjellanger et al. 2006, Goodfellow 2007, Goodfellow et al. 2007, Lidmar-Bergström et al. 2007).

It is evident that glacial erosion has altered the preglacial landscape profoundly in many areas but conversely it also seems likely that pre-Quaternary topography has conditioned glacial erosion. Pre-Quaternary fluvial valleys probably focused glacier ice flow early on in the Quaternary glacial history, thus creating a positive feedback where subsequent ice flow continued to deepen these valleys. It follows, which a large body of geomorphological literature also shows, that Quaternary glaciations have eroded the landscape selectively (e.g., Sugden 1978, Nesje and Whillans 1994, Lidmar-Bergström et al. 2000, Li et al. 2005, Fjellanger et al. 2006, Staiger et al. 2005, Phillips et al. 2006), creating mountainous areas with deeply incised troughs and virtually nonaffected uplands. These noneroded uplands are henceforth called relict nonglacial

Figure 2. Relict nonglacial surfaces east of the Lyngen fjord in Troms. Note that all mountains reach essentially the same altitude and the very flat appearance of the surfaces. These surfaces have probably been connected into a flat, undulating landscape which was uplifted and eroded through fluvial and glacial processes in the Neogene. Total relief in the area, including the fjord, is on the order of 2 km.

surfaces (Goodfellow 2007). There are two hypotheses for the preservation of relict nonglacial surfaces, 1) through thin ice where the nonglacial surfaces protrude through the ice as nunataks (e.g., Nesje and Dahl 1990) or 2) through frozen-bed preservation of the preglacial landscape beneath the ice sheet (Kleman 1992, Kleman and Stroeven 1997). The first hypothesis contends that the transition between eroded and noneroded landscape, often manifested as a trimline, reflects the ice surface, whereas the second notion maintain that the trimline reflects an ice-internal thermal boundary with possibly significant cold-based ice overriding the relict nonglacial surfaces. Applying the nunatak hypothesis creates plausible ice geometry in coastal, high-relief areas, such as along the Norwegian coast (Kverndal and Sollid 1993, Nesje et al. 2007) but further inland implies an unreasonably thin ice-sheet configuration (cf., Näslund et al. 2003, Follestad and Fredin 2011, Olsen et al. this issue). The frozen-bed preservation hypothesis, on the other hand, allows for a more realistic ice-sheet configuration in interior areas but may overestimate palaeo-ice-sheet thickness in coastal areas.

Relict nonglacial surfaces are characterised by a gently undulating topography with a relief of a few hundred metres. They are often blanketed by autochthonous block fields often showing signs of periglacial activity (cf., Goodfellow 2007, Goodfellow et al. 2007, Figure 2). Particularly on summit or ridge areas the relict nonglacial surfaces may exhibit weathering landforms such as tors (Dahl 1966, Andre 2004). Tors are delicate landforms and cannot withstand significant glacial erosion (Phillips et al. 2006). Furthermore tors probably require a long time period to develop, certainly longer than the Holocene time period, maybe even dating back to a pre-Quaternary landscape (Hättestrand and Stroeven 2002, Andre 2004, Phillips et al. 2006). Hence, tors are used as an argument for minimal glacial erosion on relict nonglacial surfaces. Also findings of soils containing clay minerals such as kaolinite and gibbsite have been invoked to infer that soils on relict nonglacial surfaces are very old and may have originated during a warmer pre-Quaternary climate, again indicating minimal glacial erosion on these surfaces (e.g., Rea et al. 1996, Marquette et al. 2004). However, it should be noted that kaolinite is difficult to distinguish from common chlorite using standard XRD practices (Moore and Reynolds 1997), furthermore both kaolinite and gibbsite have been shown to form in Holocene soils (Righi et al. 1999, Egli et al. 2001, 2004). Hence, although the gross morphology of relict nonglacial surfaces appears to be older than the Quaternary, soils draping these surfaces may well be of Quaternary age.

Relict nonglacial surfaces have also been investigated using in situ produced cosmogenic nuclides (e.g., Linge et al. 2006, Stroeven et al. 2006, Nesje et al. 2007a). In general, these investigations show that relict nonglacial surfaces have survived one or several glaciations and may even date back to pre-Quaternary times but the results are ambiguous. Recent studies suggest a multi-faceted and complex origin of blockfields (Boelhouwers 2004, Berthling and Etzelmüller 2011).

Weathering soils (saprolite), of supposed pre-Quaternary origin, have also been found at several localities in south Norway (e.g., Bergseth et al. 1980, Sørensen 1988, Olesen et al. 2007), on the Norwegian coast (Roaldset et al. 1982, Paasche et al. 2006), and in Finnmark (Olsen 1998). An overview of many published saprolite localities is given by Lidmar-Bergström et al. (1999) and Olesen et al. (2012). Again, these weathering soils have been invoked to infer limited glacial erosion also in Norway. Some of these localities have experienced little or no ice-sheet overriding such as on Andøya (Paasche et al. 2006), or they are located in sheltered positions where glacial erosion has been very limited such as in the joint-valley landscape in southern Norway (Olesen et al. 2007). However, their origin and survival through the Quaternary still remain elusive (Olesen et al. 2012).

Glacial landforms with some examples from Norway

Critical to our understanding of landscape development is the formation mechanisms of different landforms. We will here examine the most common glacial and postglacial landforms found in Norway. Glacial landforms range in size from millimetres to kilometres, i.e., six orders of magnitude (Figure 3). This span in size is also reflected in the time required for landform formation, where the smallest landforms, such as crescentic gouges may form in minutes and the largest fjords require millions of years and multiple glaciations (Figure 3). The time of formation thus range 10–12 orders of magnitude between the largest and smallest glacial landforms. We will here examine glacial landforms in Figure 3 and give some examples from Norway.

Glacial erosional landforms

Common features in all glacially scoured landscapes are micro-erosional landforms such as glacial striae, crescentic fractures and crescentic gouges. **Glacial striae** are formed when debris entrained at the sole of the glacier is dragged across a bedrock surface. Glacial striae are thus formed parallel to the ice movement and are often used to reconstruct palaeo-ice flow. Hence, glacial striae are an important element in Quaternary geological maps from the Geological Survey of Norway and other glacial geological maps (e.g., Sollid and Torp 1984). **Crescentic fractures** are up-ice convex thin arcs formed on bedrock surfaces when a debris-laden glacier sole exerts pressure on the substratum. The crescentic fractures are thus imposed on the bedrock by focused pressure formed by rocks dragged on top of the bedrock surface. **Crescentic gouges** are up-ice concave features where chips of the bedrock seem to have been removed. Smith (1984) conducted laboratory experiments with a ball-bearing pressed against a sheet of glass, thus simulating a rock sliding on top of bedrock, and

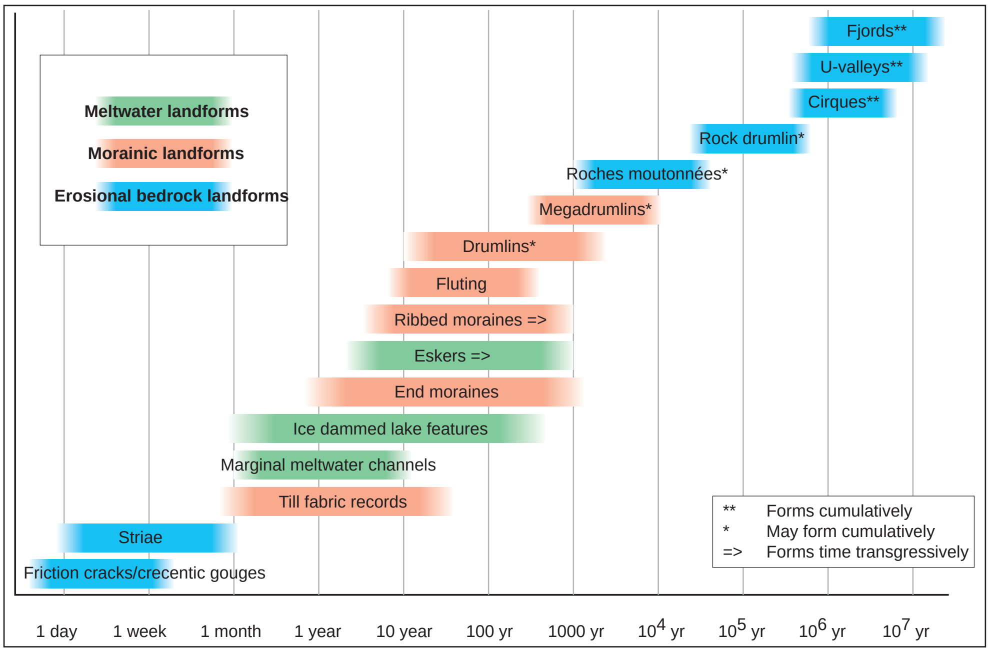


Figure 3. Glacial erosional and depositional landforms, their relative size and required time of formation. Copyright Clas Hättestrand, Stockholm University.

was able to create features similar to both crescentic fractures and gouges. Up-ice concave and convex features were shown to be conditioned by the effective pressure applied by the ball bearing during these experiments, where up-ice concave features preferably formed when a high level of pressure was applied (Smith 1984). Good examples of striae, crescentic fractures and gouges can e.g., be found in the coastal landscape of southern Norway and in front of contemporary glaciers.

Roches moutonées and **rock drumlins** are bedrock knolls that have been polished by glacial abrasion. Depending on bedrock joints and grains these landforms often become elongated in the ice-flow direction and are thus excellent indicators of former ice flow. Roches moutonées differ from rock drumlins in that they have a plucked lee-side slope, where the glacier has evacuated parts of the bedrock knoll, probably through freeze-on processes. It is debated whether roches moutonées and rock drumlins are governed by a pre-existing (pre-Quaternary) topography, perhaps akin to a stripped etch surface or if they are primarily formed by glacial abrasion (Lindström 1988, Johansson et al. 2001). Many areas in southern Norway exhibit beautiful Roches moutonées and rock drumlins; this is probably due to rapid ice movement in the run-up zone for the Norwegian channel ice stream (Sejrup et al. 2000, 2003).

Glacial cirques are large bedrock hollows with a steep headwall and an overdeepened rock-basin floor. Glacial cirques are associated with local glaciations where small cirque glaciers excavate the bedrock hollows. It is likely that many cirques are cumulative features, i.e., they are a result of multiple glaciations (Figure 3), since current cirque glaciers exhibit relatively low erosion rates and cannot likely excavate their bedrock depressions during, say, one interstadial or interglacial period (cf., Larsen and Mangerud 1981). It is also possible that full-scale ice sheets may contribute to the shaping of cirques through subglacial erosion (Holmlund 1991, Hooke 1991).

Glacial troughs are often called **U-shaped valleys** due to their characteristic cross-sectional form. These deep, linear, glacial erosional features are carved into bedrock and may require hundreds of thousands of years and several glaciations to form (Figures 3 and 4). U-shaped valleys are generally thought to have been formed mainly through glacial plucking because this process is orders of magnitude more efficient at eroding bedrock than abrasion. Since glacial erosion is dependent on ice thickness, most of the erosion takes place at the bottom of the valley, thus creating steep valley walls. The U-shaped cross profile is not properly described as 'U-shaped' but may be approximated by power-law equations or quadratic equations (Li et al. 2001, Jaboyedoff and Derron 2005). It has been shown that rock-mass strength strongly influences the shape of the cross-sectional profile, with more competent rocks causing narrower and steeper valleys (Augustinus 1992, 1995). The longitudinal profiles of U-shaped valleys commonly exhibit numerous basins or overdeepenings due to glacial plucking combined with glacier-dynamical effects such as confluence of tributary glaciers (MacGregor et al. 2000, Anderson et al. 2006). As mentioned earlier, U-shaped valleys in Norway and in many other formerly glaciated areas of the world are incised into relict nonglacial surfaces. It is generally thought that glacial troughs have developed in former fluvial (V-shaped) valleys already existing in the relict nonglacial surface; hence Quaternary glaciations have exploited and deepened pre-existing valleys. It is also likely that glacial erosion has exploited faults and weakness zones and glacial troughs often follow fault zones (Gabrielsen et al. 2002). Beautiful U-shaped valleys are found in abundance in the glacial, alpine landscape of Norway (Figure 4).

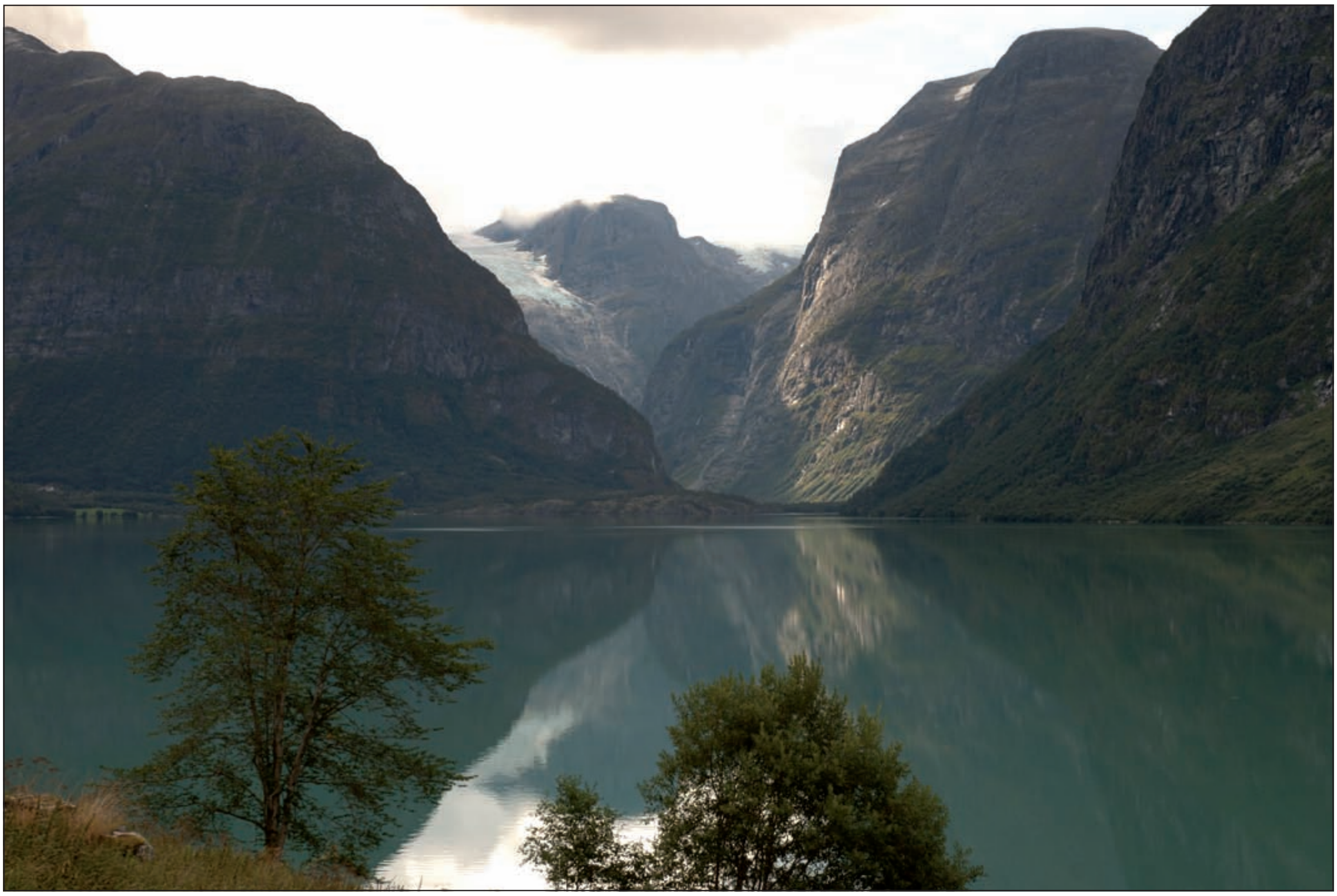


Figure 4. U-shaped valley with overdeepened lake basin at lake Loen close to Nordfjord, western Norway.

Fjords are essentially glacial troughs cut into bedrock below today's sea level and is perhaps the most prominent landscape feature in Norway (Figure 4). Fjords are formed much in the same way as U-shaped valleys although the overdeepening in fjords may be even more pronounced forming sills at the fjord mouth (Lyså et al. 2009). It is thought that these sills were formed due to less glacial erosion at the coast where the former ice tongues spread out and became thinner (Aarseth 1997). As with U-shaped valleys, fjords often follows fault zones, which may give rise to distinct fjord networks as is evident for example on the Møre coast (Gabrielsen et al. 2002). Fjords are important sinks for interglacial sediments (Aarseth 1997, Lyså et al. 2009). Indeed Aarseth (1997), calculated that about 150 km3 Holocene sediments reside in the main Norwegian fjords. Aarseth (1997) also considered most fjord sediments to be evacuated during main glaciations and, consequently, transported and deposited onto the continental shelf or shelf break.

Glaciofluvial erosional landforms

Glacial meltwater channels may either form subglacially or at the ice margin and they might be cut into bedrock or sediments. Indirectly, glacial meltwater may also cause erosion some distance from the actual ice margin. Glacial meltwater channels range in width between less than a metre up to several hundreds of metres for large canyons (Figures 3 and 5).

Subglacial channels may be formed by highly pressurised water, which is governed by the hydraulic potential gradient within the glacier. The subglacial water may thus defy gravity and cause subglacial meltwater channels that are at odds with topography or with an irregular longitudinal profile. Subglacial meltwater channels, naturally, require flowing water beneath the ice and thus indicate warm-based ice.

Marginal meltwater channels are formed by glacial meltwater flowing along the (lateral) margin of a glacier, frequently in the ablation area of the glacier where there is an abundance of meltwater and the glacier cross-profile is convex, thus diverting meltwater to the glacier margins. Marginal meltwater channels may have complete channel cross-profiles or may consist of channel floor and only one channel wall forming 'half channels'; the other channel wall having been formed on the former glacier. Marginal meltwater channels stop and end abruptly since they may meander on and off the glacier tongue or they may be suddenly diverted downwards, into the glacier through a crevasse or moulin. Marginal meltwater channels commonly form where the surface of the glacier is below the melting point; otherwise, glacial meltwater is predominantly diverted down into the glacier through percolation and melting of sinkholes. These channels are thus a relatively good indicator of melting of a cold-based ice (Dyke 1993). Extensive sets of meltwater channels are generally found in interior areas of Norway, for example highland areas of Oppland and on the

Varanger peninsula in Finnmark. Locally, different cross-cutting generations of meltwater channels can be mapped, which show overriding of nonerosive glacial ice, and facilitate reconstruction of the paleo ice-sheet configuration (Figure 5).

Impressive canyons may be cut into bedrock either by proglacial streams or more importantly by drainage of ice-dammed lakes. The famous Norwegian canyon Jutulhogget probably formed through the catastrophic drainage of the vast 'Nedre Glåmsjø' ice dammed lake about 10,300 years ago (Longva and Toresen 1991).

Smaller glaciofluvial erosional landforms include potholes (Figure 6) and P-forms. These are local features that are thought to represent highly dynamical, subglacial conditions where large volumes of meltwater are involved (Dahl 1965).

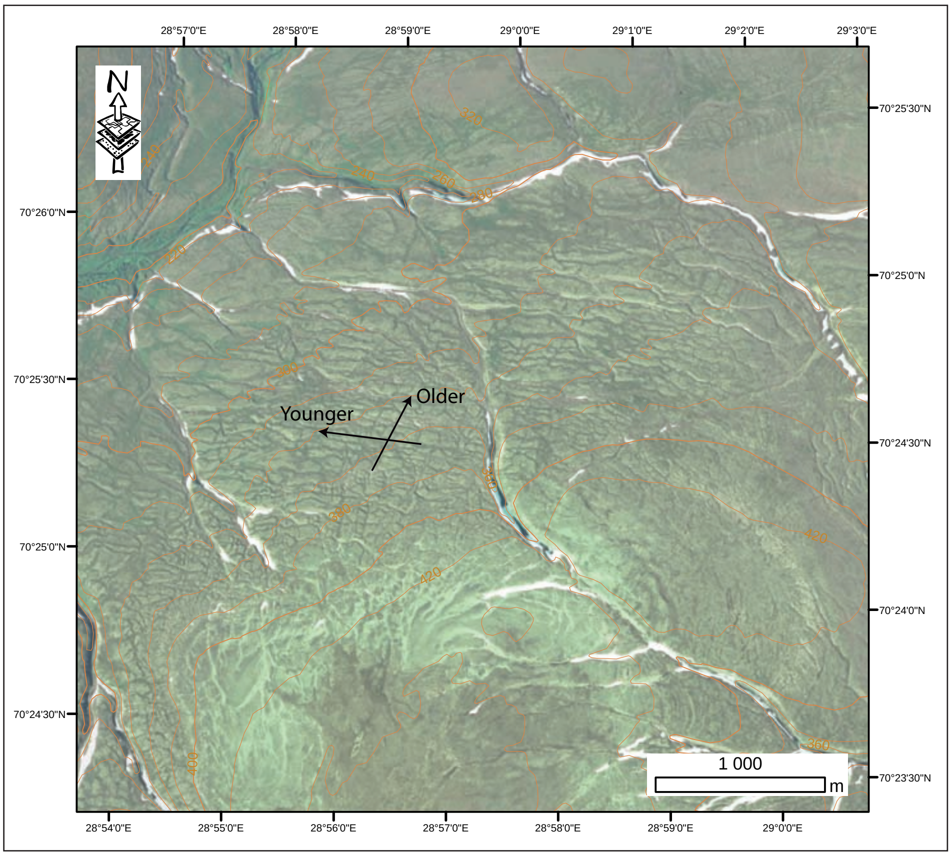


Figure 5. Aerial photograph of glacial meltwater channels on a mountain slope on the Varanger peninsula. Oldest channels are indicative of ice flow towards NNE and youngest channels indicate ice flow towards WNW (Fjellanger, 2006).

Figure 6. Large, waterfilled pothole in the upper reaches of Romsdalen valley.

Glacial depositional landforms with examples from Norway

Glacial depositional landforms

End moraines or more generally **ice-marginal moraines** are formed at the glacier margin by either: 1) glacier dumping of debris, 2) ablation or freeze-out of debris and, 3) glaciotectonic processes including pushing of proglacial sediments. Other processes might also be involved, such as rockfall or slope processes depositing material at the glacier margin. A vast amount of literature describes the various types of ice-marginal moraines and depositional processes (e.g., Clayton and Moran 1974, Boulton and Eyles 1979, Hambrey and Huddart 1995, Bennet 2001). End moraines are important glacial landforms since they indicate a glacier advance or still-stand and are thus diagnostic in reconstructing ice-sheet dynamics and configurations. One Norwegian end moraine, the Vassryggen Younger Dryas end moraine southeast of Stavanger, was used by Jens Esmark to support the theory of ice ages already in 1824 (Worsley 2006, Figure 7). Ice-marginal moraines are common in Norway and are commonly found in the proglacial areas of present glaciers. Many of these end moraines were formed during the 'Little Ice Age' (maximum at about AD 1750) and subsequent glacier retreat (Nesje et al. 1991, Mathews 2005, Burki et al. 2009). Also other Holocene end-moraine zones are common (e.g., Nesje and Kvamme 1991, Figure 8) but the most prominent ice-marginal deposits are from the Younger Dryas readvance (Andersen et al. 1995, Olsen et al. this volume). The Younger Dryas deposits commonly consist of combined sequences of end moraines and extensive proglacial outwash. Beautiful successions of Younger Dryas moraine ridges are e.g., found in South Norway (Andersen 1979), the Trondheimsfjorden area (Andersen 1979, Rise et al. 2006), and Finnmark (Olsen et al. 1996, Sollid et al. 1973).

Hummocky moraine is characterised by irregular morphology and sediment structure. It is typically formed at or on glaciers with a high content of debris (Gravenor and Kupsch 1959, Eyles 1979) and, if widespread, may represent stagnant disintegration of the ice sheet. Examples of areas with hummocky moraine may be found on Jæren (Knudsen et al. 2006) and on Finnmarksvidda. **Ribbed moraine**, which sometimes is classified as a type of hummocky moraine, is very widespread in northern Norway and at some altitude or interior areas of Norway (Sollid and Sørbel 1994). Ribbed moraine may also be called Rogen moraine (Figure 9), from the type locality at lake Rogen on the Swedish–Norwegian border. Ribbed moraines are transverse moraine ridges formed subglacially, which is apparent from the fact that they often have drumlins superimposed on them. They commonly have a somewhat arcuate plan form with the concave elements oriented down-ice. There are a number of hypotheses on the formation of ribbed moraine, including subglacial water causing megascale ripples, squeeze-up of subglacial till into crevasses and glaciotectonic stacking of till slabs (Bouchard et

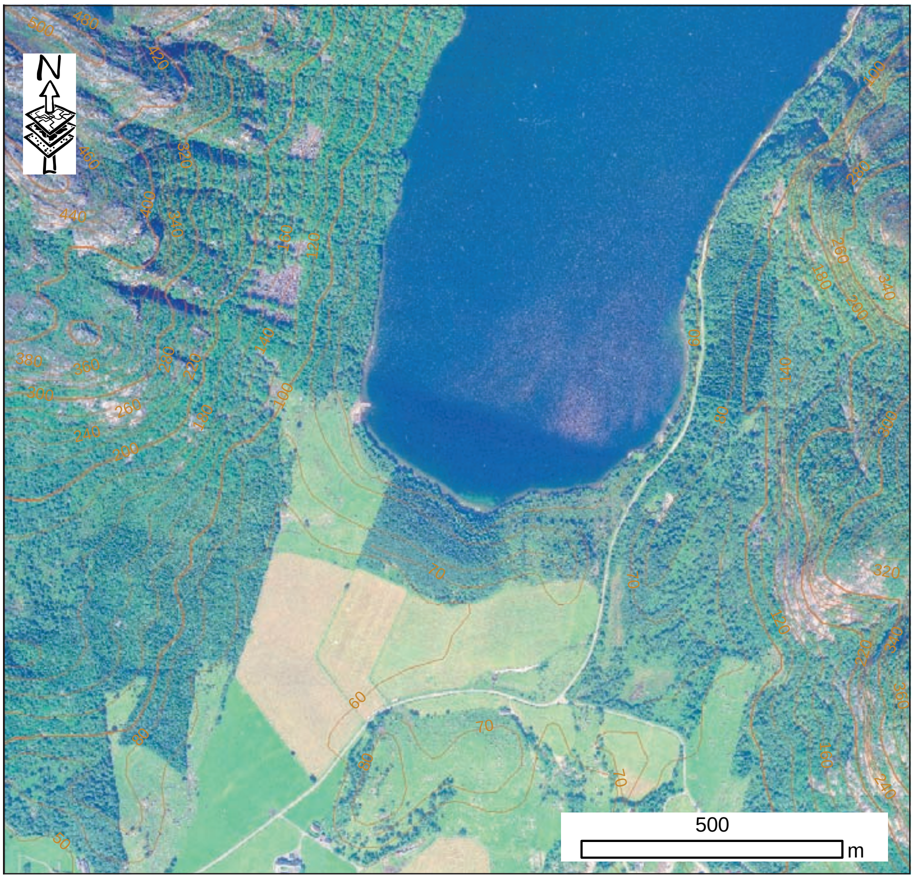


Figure 7. Aerial photograph of the Vassrygg Younger Dryas end moraine and proglacial outwash, situated just south of the Haukalivatnet.

Figure 8. The Ølfjellet moraine in Nordland of likely Pre-Boreal age. Note multiple and complex ridges indicating an oscillating ice margin.

Figure 9. Rogen moraine at Hartkjølen in Nord-Trøndelag.

al. 1989, Hättestrand 1997). Based on extensive mapping of ribbed moraines Kleman and Hättestrand (1999) argued that ribbed moraine forms at the transition between cold-based ice and warm-based ice, which is also indicated by Sollid and Sørbel (1994). Ribbed moraine might thus be a good indication of areas beneath an ice sheet where basal temperatures have been low. Extensive areas of ribbed moraine may be found in northernmost and eastern Norway (Sollid and Torp 1984, Sollid and Sørbel 1994) and many smaller areas can be found in elevated terrain such as in Jotunheimen.

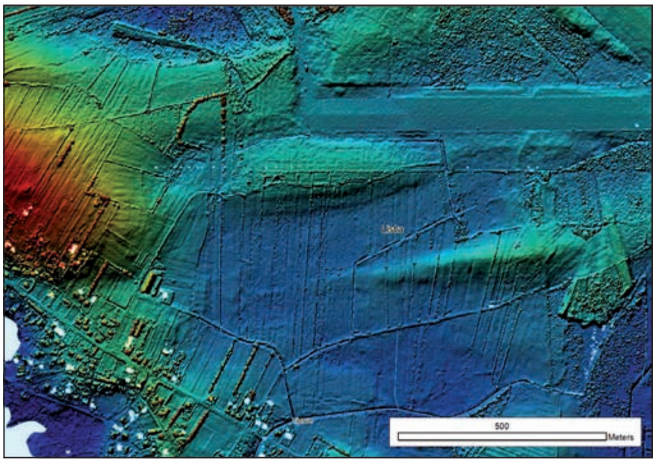


Figure 10. Drumlins on Lista in South Norway shown in a shaded relief map produced through high-resolution LiDAR digital-elevation data. The drumlins show ice flow from ENE towards the Skagerrak ice stream.

Flutes and drumlins are streamlined ridges composed mainly of glacial sediments with a subglacial genesis. Glacial flutes are small (<3 m high and <3 m wide) and are commonly found in front of current glaciers. Their small size makes them likely to decay and disappear over longer time periods, and hence they are less distinct in areas deglaciated some time ago. Flutes occuring as erosional forms in bedrock (rock flutes) may occasionally be found, e.g., at the head of Glomfjord in Nordland county (Gjelle et al. 1995). Drumlins are larger, typically ranging between 5 and 50 m in height and may be a few kilometres long; in addition they have a length-to-width ratio of less than 50 (Bennet and Glasser 1996). Drumlins commonly exhibit a typical shape with the blunter end pointing in the up-ice direction and the elongated tail pointing in the down-ice direction; they are thus excellent indicators of former ice-flow direction (Kleman and Borgström 1996). If drumlins are superimposed on each other they give a direct evidence, and relative chronology, of shifting ice-flow patterns (Clark 1993). Formation of drumlins is not fully understood although most researchers argue that subglacial deformation, possibly conditioned by subglacial obstacles, is the main process involved in the building of drumlins and related landforms. Beautiful drumlin fields in Norway can e.g., be found northeast of Dombås (Follestad and Fredin 2007), Lista in South Norway (Figure 10) and on Finnmarksvidda (Figure 11).

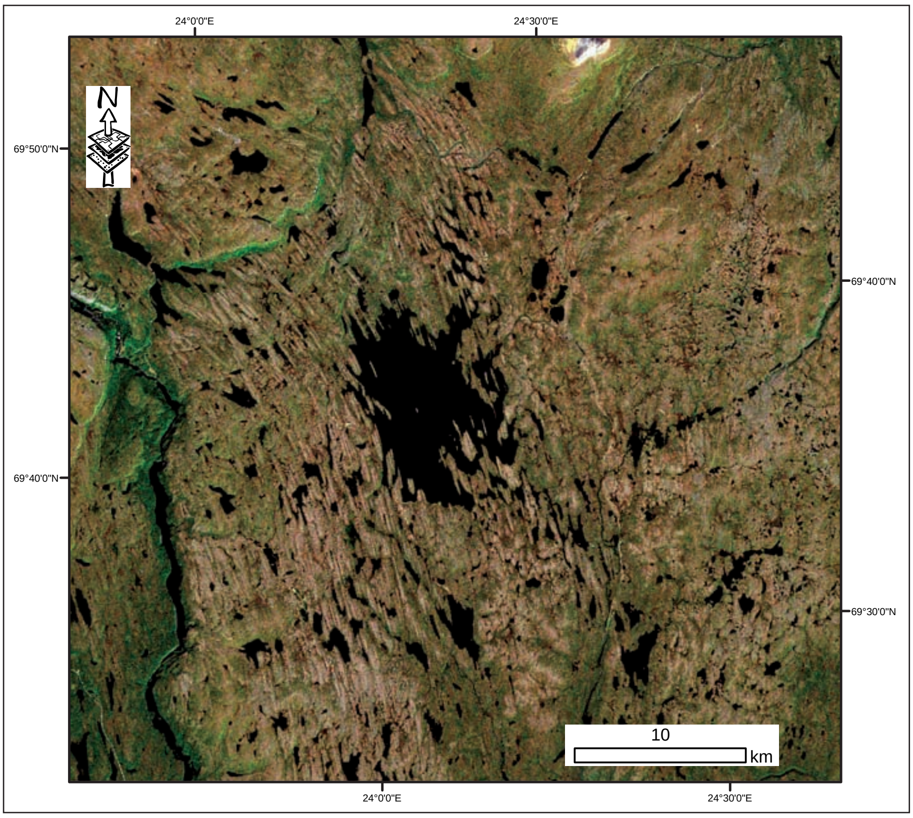


Figure 11. An extensive drumlin swarm on Finnmarksvidda, northern Norway, depicted in a Landsat 7 satellite scene. The drumlin swarm 'bends' due to migration of the ice-dispersal centre during the last deglaciation, thus showing the time-transgressive nature of drumlin formation. Also note the esker running from south towards north to the right in the picture.

Glaciofluvial depositional landforms

Eskers consist of sand, gravel and cobbles, and are formed in tunnels beneath the glacier by glacial meltwater (Brennand 2004). They are generally thought to form time transgressively, relatively close to the ice margin. Eskers can in rare cases be several hundreds of kilometres long and contain significant volumes of sediments, which are valuable resources for extracting groundwater or gravel. Since eskers form obliquely to the glacier margin, towards the ice front or downslope from the lateral ice margin, they are very useful for reconstructing glacier-margin retreat. Eskers may be found in many areas in Norway, and generally reflect the Weichselian deglaciation pattern. Long and continuous eskers can easily be traced in Finnmark (Figure 11).

Ice-dammed lake features such as perched glaciofluvial deltas are formed at or close to an ice margin when glacial meltwater enters a water body, which in turn often is glacially dammed. Glaciofluvial deltas often exhibit palaeo-water channels on the top surface, different levels of sedimentation if the glacial lake has had changing lake levels, and may be cut and eroded by postglacial fluvial erosion (Figure 12). Glaciofluvial deltas are important indicators of glacial-lake dynamics, including damming and glacial-lake bursting. Good examples of glaciofluvial deltas are found in south-central Norway such as the Dørålsæter or Grimsmoen delta (Figure 12). Widespread and complex fine-grained sediments can be found in south-eastern parts of Norway. Many of these deposits were laid down in the glacial lake Nedre Glåmsjø, which was drained about 10,300 years ago (Longva and Bakkejord 1990, Longva and Thoresen 1991, Berthling and Sollid 1999).

Glacial landforms through the last glacial cycle

Several attempts have been made to link regional glacial landform sets to different time periods during the last glacial cycle (e.g., Kleman et al. 1997, Boulton et al. 2001). These two reconstructions have later been used to validate numerical ice-sheet models with favourable results (Näslund et al. 2003). This type of work relies on extensive mapping of glacial landforms, mainly drumlins, moraine ridges, eskers and glacial meltwater channels. Subsequent to mapping it is important with data reduction and assembling the landforms into coherent sets of ice-flow patterns. In addition, it is of key importance to connect these landforms into both morphostratigraphic and absolute chronologies.

The Early and Middle Weichselian (marine isotope stages 5–3)

During marine isotope stages 5 through 3, the Weichselian ice sheet was predominately centred in the Scandian mountain range (Porter 1989, Kleman et al. 1997, Boulton et al. 2001, Fredin 2002). Large-scale, erosive landforms, such as fjords and glacial through valleys, probably continued to develop during this time period (Porter 1989, Fredin 2002).

Landforms from the Early and Middle Weichselian (marine isotope stages 5–3) are rare, although several localities from many parts of Norway with Early to Middle Weichselian sediments have been reported (overviews by Mangerud 2004, Mangerud et al. 2011, Olsen et al. this volume). Much of the landform record from these stages has probably been obliterated in Norway by subsequent glacial stages, although widespread glacial-landform systems of Early Weichselian age have been reported from Sweden and Finland (e.g., Lagerbäck 1998, Hirvas 1991, Kleman 1992, Hättestrand 1998, Fredin and Hättestrand 2002).

Glacial striae reported by Vorren (1977, 1979) and Sollid and Torp (1984) in South Norway are perhaps the only likely

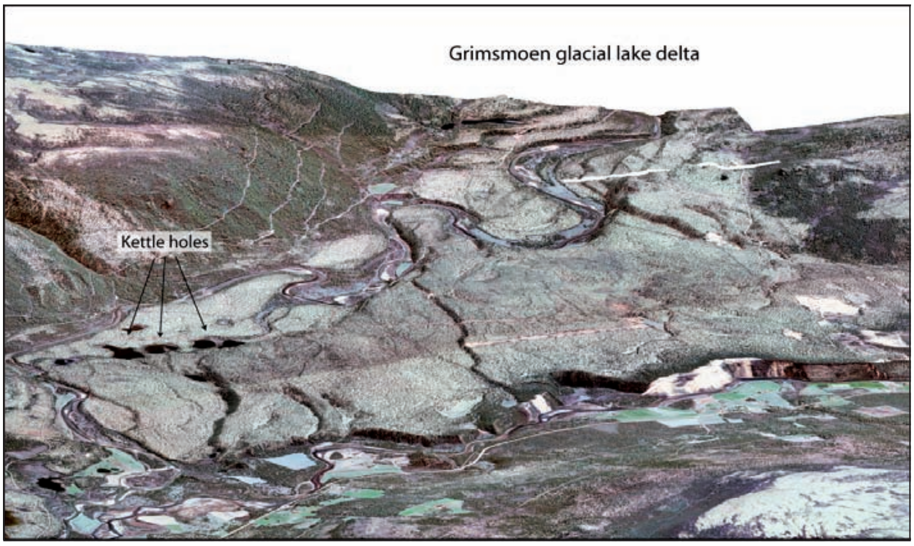


Figure 12. Aerial photographs draped over a digital elevation model (topography) showing a 3D view of the Grimsmoen glacial-lake delta in south-central Norway. Note the different terrace levels indicating lowering of the glacial lake during deposition. Kettle holes are evidence for stranded icebergs in the delta. Holocene river erosion has cut into the frontal parts of the delta.

candidates for an Early Weichselian ice-flow pattern emanating from the upland areas around Hardangervidda, as indicated by Kleman et al. (1997). This is in accordance with the general notion that ice-sheet initiation centres probably were situated in the highest areas around Jotunheimen/Hardangervidda in the south and the Sulitjelma/Sarek/Kebnekaise massifs in northern Sweden and Norway (Kleman et al. 1997, Boulton et al. 2001, Fredin 2002).

On the Varanger peninsula, northern Norway, there are conspicuous sets of lateral meltwater channels that are clearly older than the last glacial maximum. These meltwater channels are only dated relative to younger, cross-cutting meltwater channels of supposed Late Weichselian age (Fjellanger 2006). Also, locally in the same area, circular ablation-moraine mounds are superimposed on the older meltwater channels, again indicating a pre-LGM (Last Glacial Maximum) origin for these channels (Ebert and Kleman 2004, Fjellanger 2006). It is not possible to say during which time these meltwater channels formed since no absolute dates exist, but based on geomorphological criteria it is clear that they were created during a frozen-bed deglaciation (cf., Dyke 1993, Kleman and Borgström 1996).

In interior areas in southeast Norway there are numerous sets of end moraines, drumlins, Rogen moraine and glacial meltwater channels that likely predate the last glacial maximum (Sollid and Sørbel 1994, Fredin 2004). Nice examples of supposedly pre-LGM meltwater channels and marginal moraines are for example found on and around the Stølen mountains close to lake Femunden. This general supposition is based on the observation that these landforms are generally incompatible with known LGM and deglaciation ice-flow patterns in the area and considerations of the thermal regime in the ice (Sollid and Sørbel 1994, Fredin 2004).

The last glacial maximum and deglaciation (marine isotope stages 2–1)

The vast majority of glacial landforms in Norway probably date to the last glacial maximum or the deglaciation. During the build-up towards the last glacial maximum, the main Weichselian ice-dispersal centre migrated towards the east into central Sweden in the northern sector, and remained in the interior areas in south-central Norway (Kleman et al. 1997, Mangerud et al. 2011). Ice streams radiated from the large ice sheet and flowed in the larger larger valleys through the mountain range, focused into fjords and continued onto the continental shelf (Dowdeswell et al. 2006). Ottesen et al. (2009) give an overview on glaciations on the Mid-Norwegian continental shelf. It is likely that large-scale erosion of mountainous valleys and fjords continued and intensified during this time period. In addition low-lying, flatter areas, such as on the south coast of Norway, were subjected to aerial scouring in the run-up zone for the massive Norwegian Channel ice stream (Sejrup et al. 2003). Although the ice sheet at its maximum probably was about 3 km thick in its core, peripheral positions such as along the coast were ice free, at least in higher terrain (e.g., Nesje et al. 1987, Nesje et al. 2007, Olsen et al. this volume). It has also been suggested that lower terrain in interior areas of Norway might have been be ice free during this time period (Nesje and Dahl 1990, Paus et al. 2006). This is refuted by Olsen et al. (this issue), who show that the ice surface must have been at a higher elevation. Following the glacial maximum, the ice sheet retreated across the continental shelf onto the coast until the Younger Dryas still-stand/readvance at about 12900–11500 BP. During this period, large ice-marginal deposits were formed including both morainic landforms and glaciofluvial outwash (Andersen et al. 1995). During the subsequent deglaciation, glacial landforms continued to develop, e.g., drumlins and eskers around Dombås (Follestad and Fredin 2007) and on Finnmarksvidda (Figure

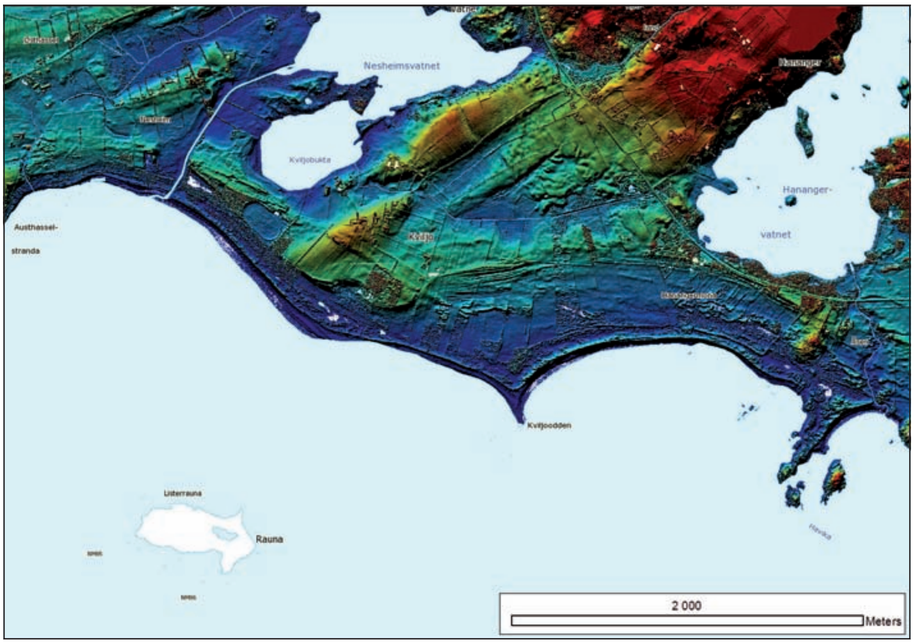


Figure 13. Shaded relief map showing large drumlins and multiple fossil shorelines on Lista, southernmost Norway. Note the sequence of shorelines between Kviljo and Hanangermona. The map is produced from high-resolution LiDAR digital elevation data.

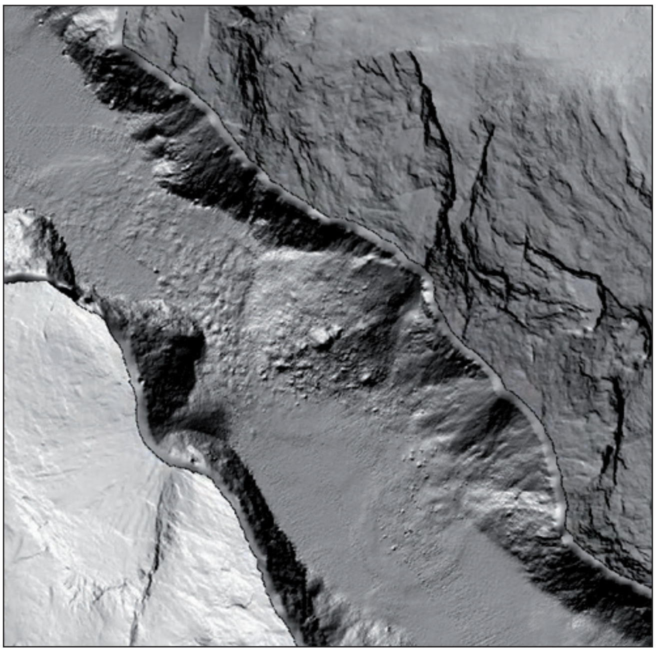


Figure 14. Shaded relief in a combined bathymetric and topographic dataset showing the floor of Tafjord. Note the very large submarine rock-avalanche deposits in the fjord.

11). Because the main ice divide in South Norway was situated south of the water divide, large ice-dammed lakes formed, at least partly subglacially–sublaterally, during the deglaciation leading to deposition of ice-dammed lake deposits. When these lakes drained, extensive meltwater erosional landforms were formed together with erosional marks from floating icebergs (Longva and Thoresen 1991).

Minor readvances into the final deglaciation led to the formation of still more ice-marginal and glaciomarine deposits at many localities in Norway (e.g., Nesje et al. 1991). Furthermore, fjord and lacustrine systems received large amounts of glacial and glaciofluvial sediments during the deglaciation, forming vast deposits (Aarseth 1997).

Postglacial landforms—the Holocene landscape development

Shortly after the last deglaciation, reworking and redistribution of glacial landforms and sediments started mainly through the action of fluvial, gravitational, and coastal processes (Ballantyne 2002). There have recently been significant advances in the understanding of sediment budgets in deglaciated catchments (Ballantyne and Benn 1994, Beylich et al. 2009, Hansen et al. 2009, Burki et al. 2010).

In coastal areas much of the postglacial landform development is conditioned by the former ice sheet through the isostatic rebound and associated shoreline displacement in combination with wave action (Figure 13). Furthermore, fluvial processes during the Holocene has eroded, transported and deposited vast amounts of sediment, typically derived from glacial landforms. Finally, gravitational processes affect both steep slopes, e.g., in the form of rock slides (Figure 14) and debris flows (Figure 15). Graviational processes also act on glaciomarine clay deposits through quick-clay slides (Figures 16 and 17).

Marine landforms

Shorelines are usually erosional landforms indicating wave action for a prolonged time period, and are thus indicative of a sustained sea-level stand. On the Norwegian coast, shorelines are common and the most pronounced shorelines typically reflect the highest sea-level stand following the deglaciation, the Younger Dryas 'main line', or the Tapes transgression shoreline at around 6300 BP (cf., Andersen 1968, Svendsen and Mangerud 1987, Sørensen et al. 1987, Reite et al. 1999, Romundset et al. 2011). Shorelines may also be formed at the shores of glacial lakes and perched shorelines may be found in abandoned glacial-lake basins. Somewhat related to shorelines are **coastal caves**, which are also created by coastal erosion (Figure 18). They typically form where a lithological weakness zone in coastal cliffs coincides in altitude with a prolonged sea-level stand, thus allowing wave erosion to act on the weakness zone. Coastal caves are found, e.g., on the Møre coast and probably date to the Eemian high sea-level stand (Larsen and Mangerud 1989).

Figure 15. Debris-flow scars and deposits in Ulvådalen west-central Norway. Many debris flows were triggered in Ulvådalen during the spring melt of 1960.

Figure 16. A landscape at Ulset dominated by glaciomarine clays with significant quick-clay slide scars.

Perched glaciomarine deltas or deposits are widespread phenomena in Norway and reflect deposition of glaciofluvial material into the sea directly after deglaciaton (Eilertsen et al. 2005). These deposits are very important for reconstruction of the local marine limits since the top surface, or more precisely the contact between the topset and the foreset layers of the delta, reflects the sea-level stand at the time of deposition (Figure 19).

The **strandflat** is a flat bedrock platform along most of the Norwegian west coast, which extends both above and below the present sea-level stand. It is close to 60 km wide at the Helgeland coast and absent in other coastal areas e.g., at Stad. Also, strandflat-like features, or strand terraces, at inland lakes have been reported by e.g., Aarseth and Fossen (2004). A wide range of processes accounting for development of the strandflat has been proposed. Reusch (1894) argued that marine abrasion (wave action) was the main agent for developing the strandflat. Erosion by sea ice and frost weathering was proposed by Larsen and Holtedahl (1985), which is an explanation also favoured at

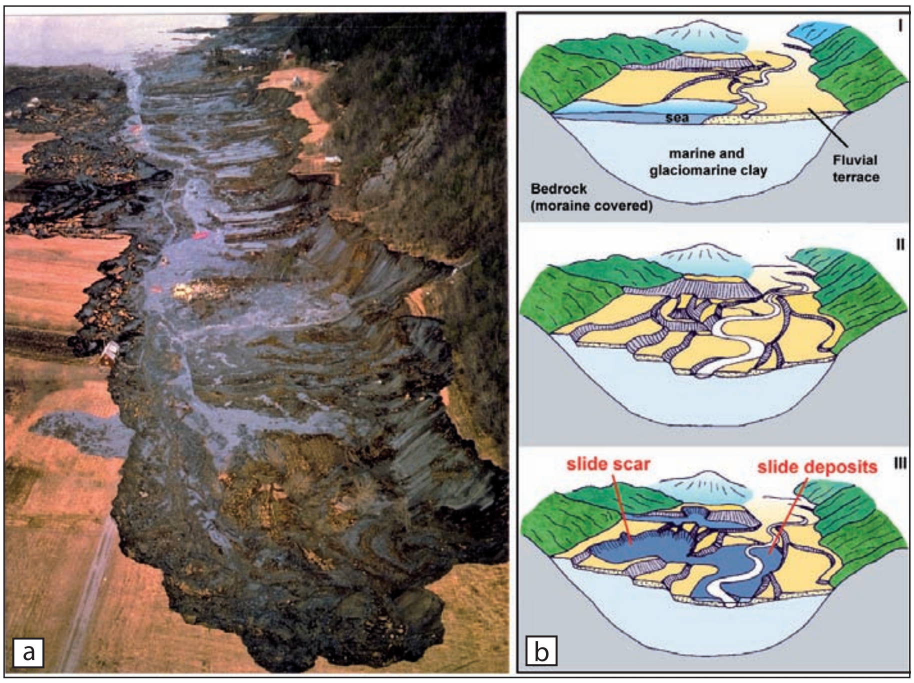


Figure 17. (a) Press photograph of a quick-clay slide at Rissa, mid-Norway. (b) Cartoon showing the evolution of quick-clay slides. Shortly after deglaciation isostatic rebound lifts the landscape causing fluvial incision, which in turn might trigger quick-clay slides.

Figure 18. The cave Harbakhula (left-central in the picture) in the coastal areas of Trøndelag. The cave was formed through marine abrasion during high sea-level stands in the Quaternary.

Figure 19. The Fremo glaciomarine deltaic terrace south of Trondheim. The top surface of the terrace reflects the sea-level stand directly after deglaciation. Note the pronounced sloping beds (foresets), which prograded out into the sea during delta deposition.

the lacustrine strand terrace described by Aarseth and Fossen (2005). Dahl (1947), on the other hand, argued that glacial erosion was an important factor in shaping the strandflat. The strandflat can slope gently out from the coast or be essentially horizontal. Holtedahl (1998) argued that the sloping strandflat was forming at or close to the isostatically migrating shoreline following a deglaciation, while the horizontal strandflat represents a more mature stage in the strandflat development.

Quick-clay slides and scars are formed where glaciomarine clay was deposited shortly after the deglaciation and where isostasy subsequently has lifted up the landscape (Figure 17). Rain and groundwater flow has washed out the salts in the clay deposits during the Holocene, causing alteration of the intra-grain structure, which in turn destabilises the clay deposits. Coastal areas in Norway below the marine limit is thus subject to quick-clay slides. Numerous quick-clay slide scars in, e.g., mid-Norway (Reite et al. 1999) northern Norway (Hansen et al. 2007) provide evidence for extensive quick-clay slide activity throughout the Holocene.

Gravitational processes

Rock slides, rock avalances and falls are primarily conditioned by structural geology (bedrock competence and structure) and the lingering effects of Quaternary glaciations. Glaciations control slope stability in three ways. First, glacier erosion will cause increased relief and steepening of valley slopes, which in turn

Figure 20. Photograph of a potential rock avalanche on Nornesfjellet, threatening to fall into the Lyngen fjord and thus jeopardising several communities in the area. There are several similar sites in the area. All are situated on relatively steep glacial trough walls in a zone of high seismic activity (Iain Henderson, pers. comm.).

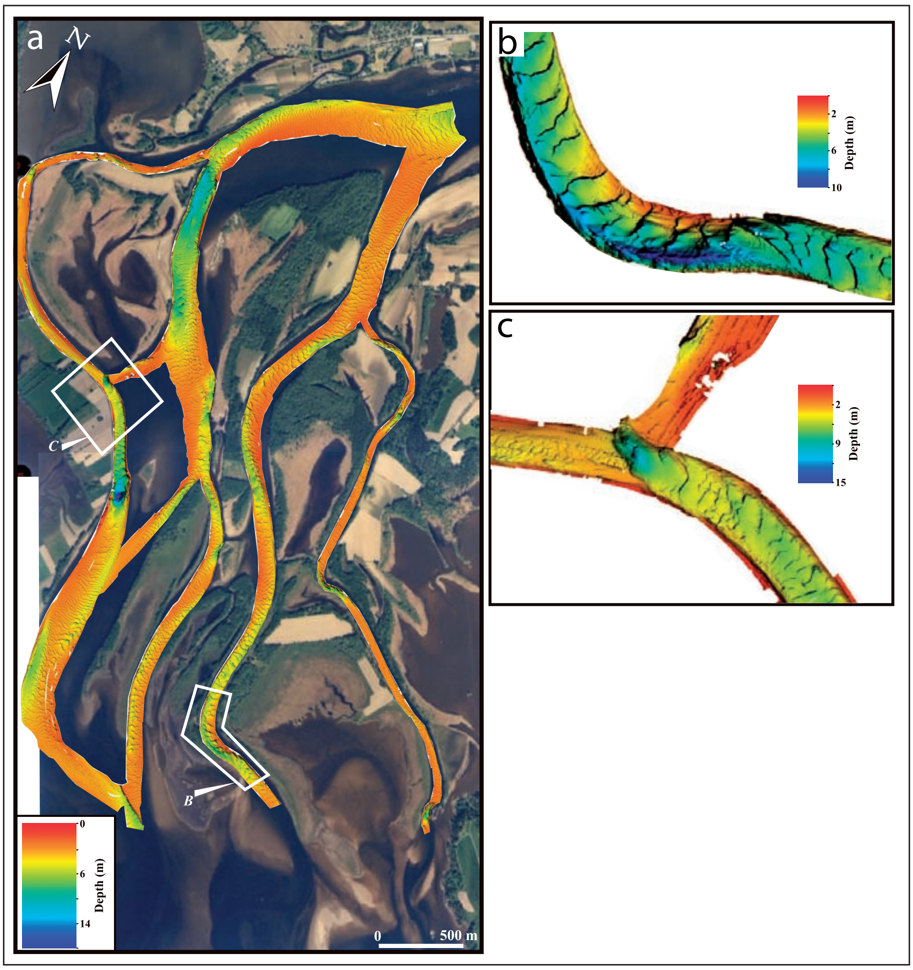
will cause increased overburden stresses in the rock mass. Second, deglaciation (glacier melting) of a landscape causes debuttressing when the supporting ice melts. Unloading of glacier ice may thus cause release of stress, which was built up in the rock mass during glacial loading (Augustinus 1992, 1995, 1996, Ballantyne 2002). Third, landscape-scale isostatic rebound following ice-sheet melting may cause significant postglacial earthquakes and faulting. This seismic activity may in turn trigger rock-slope failure in already susceptible areas (Lagerbäck 1990). Seismic activity of other origins seems also to be instrumental in triggering rock slides and geohazards (Blikra 1999, Blikra et al. 2000). Ballantyne (2002) argues that rock-fall activity should be highest shortly after the main deglaciation, but evidence showing that all types of seismic activity may trigger rock avalanches indicates that this is an ongoing process. Indeed, several localities in Troms (Figure 20) and Møre & Romsdal are currently monitored to mitigate rock-avalanche hazards. The biggest risk associated with these rock avalanches is that they might fall down into a fjord and cause a local surge—tsunami—that might threaten lives and infrastructure. Bathymetric investigations have shown that this has happened repeatedly in Norwegian fjords throughout the Holocene (Figure 14).

Debris flow and other slope processes where water is involved is also a very common process in Norwegian alpine terrain (Figure 15). Ballantyne (2002) argues that debris-flow activity, and other slope processes, was highest closely following the last deglaciation due to unstable sediments and general landscape readjustment. This is to some degree confirmed by studies in Norway, but also other periods with high slope-process activity during the Holocene have been detected, possibly linked to climate variations (e.g., Blikra and Nemec 1998, Sletten et al. 2003, Sletten and Blikra 2007).

Fluvial processes

Holocene fluvial erosion is, perhaps, the main agent reshaping the postglacial landscape (Church and Ryder 1972, Ballantyne 2002, Eilertsen 2002). Glacial depositional landforms are very susceptible to fluvial erosion shortly after the deglaciation due to little vegetation and unstable sediment structures (Ballantyne 2002). Although sediment stability increases over time, as vegetation stabilises sediments and the deposits generally adapt to the new environment, fluvial erosion continued throughout the Holocene. These sediments are then transported in rivers, of ranging magnitude, into sinks in the form of lakes or the sea. Transport mechanisms include movement of very coarse-grained material in steep, alpine catchments to fine-grained, slow transport in low-gradient catchments and rivers. Where these water pathways enter a lake or the sea, deltas are often formed such as where Glomma enters Lake Øyeren in South Norway (Figure 21).

Figure 21. (a) Bathymetry of channels in the river Glomma delta where it enters Lake Øyeren in South Norway. (b) Bathymetry showing bottom structures (transverse bars) in a meander exhibiting the typical deeper channel along the outer bank. (c) Bottom structures at a channel intersection, note the deep pool formed at the intersection.



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Quaternary glaciations and their variations in Norway and on the Norwegian continental shelf

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In this paper our present knowledge of the glacial history of Norway is briefly reviewed. Ice sheets have grown in Scandinavia tens of times during the Quaternary, and each time starting from glaciers forming initial ice-growth centres in or not far from the Scandes (the Norwegian and Swedish mountains). During phases of maximum ice extension, the main ice centres and ice divides were located a few hundred kilometres east and southeast of the Caledonian mountain chain, and the ice margins terminated at the edge of the Norwegian continental shelf in the west, well off the coast, and into the Barents Sea in the north, east of Arkhangelsk in Northwest Russia in the east, and reached to the middle and southern parts of Germany and Poland in the south. Interglacials and interstadials with moderate to minimum glacier extensions are also briefly mentioned due to their importance as sources for dateable organic as well as inorganic material, and as biological and other climatic indicators.

Engabreen, an outlet glacier from Svartisen (Nordland, North Norway), which is the second largest of the c. 2500 modern ice caps in Norway. Present-day glaciers cover together c. 0.7 % of Norway, and this is less (ice cover) than during >90–95 % of the Quaternary Period in Norway.

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Introduction

This paper is one of a series of three papers presenting the Quaternary[1] geology of Norway and, briefly, the adjacent sea-bed areas. The utilised information comes from various sources, but focuses on data from NGU where this has been possible (NGU-participation in research). In a few cases, both from land and sea-bed areas this has not been possible and external sources have been included.

This paper concerns both data from mainland Norway and from the continental shelf, both for the pre-Weichselian and Weichselian history. The reconstruction of the Last Glacial Maximum (LGM) interval and late-glacial ice-sheet fluctuations will be discussed more thoroughly.

The temperate to warm intervals of interglacial and interstadial status have occupied more than half of the Quaternary, but the glacial intervals are in focus here (Figure 1a–d).

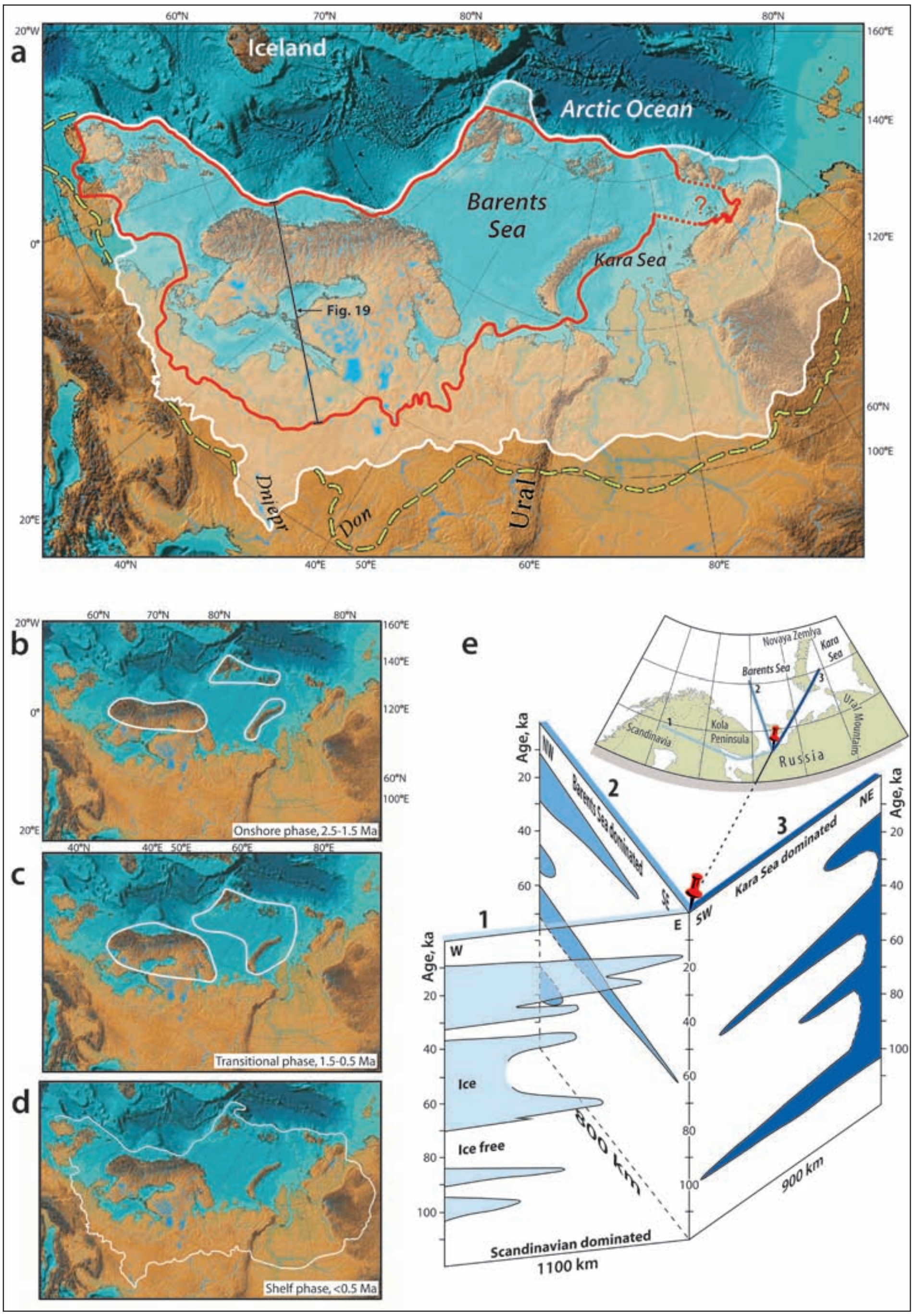


*Figure 1. Glaciations/glacial extension in northern Europe. (a) LGM (*red line*) and previous major glaciations (*white line*: Saalian and Elsterian glaciations;* yellow stippled line: *Quaternary maximum glacier extension), with coalescence between the Scandinavian–Fennoscandian, Barents Sea–Kara Sea and British Isles ice sheets. Modified from Svendsen et al. (2004). (b), (c) and (d) Style of maximum type of glaciations within three periods of the Pleistocene (with redefined lower boundary at c. 2.6 Ma). The onshore and transitional phases are conceptual. After Larsen et al. (2005). (e) Weichselian glaciation curves depicting Scandinavian ice sheet dominance, Barents Sea ice sheet dominance and Kara Sea ice sheet dominance. After Larsen et al. (2005).*

[1] Quaternary is here used according to the formally redefined lower boundary at 2.6 million years before present, which now also defines the base of the Pleistocene (Gibbard et al. 2009, Mascarelli 2009) – an epoch that encompasses the most recent glaciations, during which the glaciers started to grow bigger, and much more frequently than before extended offshore as reflected in ice-rafted debris in deep-sea sediments (e.g., Bleil 1989, Jansen and Sjøholm 1991, Jansen et al. 2000).

The glacial history of Norway can be compared to those of the neighbouring formerly glaciated areas in the British Isles, the Barents Sea and the Kara Sea. As illustrated for the last glaciation from an area in the northeast (Figure 1e) the Fennoscandian/Scandinavian[2], Barents Sea and Kara Sea ice sheets have met during the LGM, but a detailed correspondence in time between glacial fluctuations in any two or all of these areas should not be expected (Larsen et al. 2005).

The glacier extension is inferred from the location of the associated till(s), other glaciogenic deposits and their associated ice-flow directions. For example a till which simply is *present* on eastern Finnmarksvidda, implies on its own a continental ice sheet that is reaching at least to the fjords of northern Fennoscandia. Furthermore, an ice-flow direction towards the NW associated with a till bed located on Finnmarksvidda or in northern Finland indicates a Fennoscandian ice sheet with an ice dome/ice-shed zone over Finland (**F** configuration) of a thick ice that moved almost topographically independent even in moderate to high-relief fjord areas. This led to a large ice extension reaching to the shelf area, and possibly to the shelf edge (Olsen et al. 1996a, and Figure 2a, b). Smaller extensions from Scandinavian or mountain-centred glaciations (**S** configuration), representing a Scandinavian ice sheet are associated with ice-flow directions towards the NNE–NE across Finnmarksvidda and northernmost parts of Finland, whereas those of medium-sized glaciations reaching to the outer coastal zone/inner shelf areas are expected to have intermediate ice-flow directions, which is towards the N across Finnmarksvidda and northernmost parts of Finland.

Previous reviews on the Quaternary glacial history of Norway and adjacent sea-bed areas (North Sea, Norwegian Sea, Barents Sea), e.g., Holtedahl (1960), Andersen (1981, 2000), Thoresen (1990), Vorren et al. (1990, 1998), Holtedahl (1993), Andersen and Borns (1994), Jørgensen et al. (1995), Sejrup et al. (1996, 2005), Mangerud et al. (1996, 2011), Mangerud (2004), Dahlgren et al. (2005), Hjelstuen et al. (2005), Nygård et al. (2005), Rise et al. (2005), and Vorren and Mangerud (2008) describe an enormous erosional impact on the Norwegian landscape, producing deep fjords and their extensions on the shelves, long U-shaped valleys, numerous cirques and many lakes in overdeepended bedrock basins. Examples and details on this are given in *"Glacial landforms and Quaternary landscape development in Norway"* (Fredin et al., 2013). More examples and details of deposits and stratigraphy, though mainly from onshore localities are presented in *"Quaternary glacial, interglacial and interstadial deposits of Norway and adjacent onshore and offshore areas"* (Olsen et al. 2013).

We refer also to Ottesen et al. (2009), which gives specific examples of the glacial impact with erosion and deposition on the Mid-Norwegian continental shelf areas. Furthermore, for the northern sea-bed areas we refer to e.g., Laberg et al. (2010) who have described the late Pliocene–Pleistocene palaeoenvironment from the southwestern Barents Sea continental margin.

In this paper ^{14}C ages younger than 21,300 ^{14}C years BP have been calibrated to calendar (cal) years BP according to INTCAL04.^{14}C (Reimer et al. 2004) and MARINE04.^{14}C (Hughen et al. 2004). To convert older ages to calendar years 4,000 years are simply added to the ^{14}C age (Olsen et al. 2001a). The abbreviation ka, meaning a thousand years, is here used as a thousand years before present, so that the BP in ka BP is generally omitted.

The oldest recorded Cenozoic glacial history in Norway

The deep-sea record (data from the Vøring Plateau) and old regional land and sea-bed data

The records of the oldest glaciations are represented by IRD (ice-rafted detritus) in deep-sea sediments (Figure 2b). A review by Mangerud et al. (1996) concluded that calving Cenozoic glaciers first occurred along the coast of Norway *c.* 11 million years ago (Ma). By comparing the sedimentary stratigraphy of the Netherlands, the global marine oxygen isotope signal, and the amount of ice rafting these authors found that there was a significant increase in the size of Fennoscandian/Scandinavian ice sheets after the onset of the Praetiglian in the Netherlands at 2.5–3 Ma (Zagwijn 1992). Sediments of Menapian age (c. 1.1 Ma) include the oldest strata in the Netherlands and North Germany that carry large quantities of Scandinavian erratics from east Fennoscandia and Central Sweden. They may be interpreted as evidence of a first major ice advance beyond the limits of the present Baltic Sea (Ehlers et al. 2004), and this ice sheet must obviously have covered most of Norway too. In comparison, the oldest documented major Barents Sea ice sheet in the southwestern Barents Sea area is supposed to have occurred almost 1.5 million years ago (Andreassen et al. 2007).

The border zone of the Scandinavian ice sheets in the west and south

The Norwegian continental shelf

The Fedje Till, which is tentatively assigned an age of *c.* 1.1 Ma and recorded in the Norwegian Channel, has until recently been regarded as the oldest identified and dated glacial deposit on the Norwegian shelf (Sejrup et al. 1995). However, new results from sediments underlying the Fedje Till indicate that several older glacial erosional horizons occur in this area, and the oldest may be as old as *c.* 2.7 Ma (A. Nygård, pers. comm. 2007). This may correspond with the base of the Naust Formation on the Mid-Norwegian continental shelf, which is considered to have a glacial origin and an age of *c.* 2.8 Ma (Ottesen et al. 2009). It may also correspond with the increased level of IRD in the

[2] Fennoscandia = Finland, Sweden and Norway; Scandinavia = Sweden and Norway.

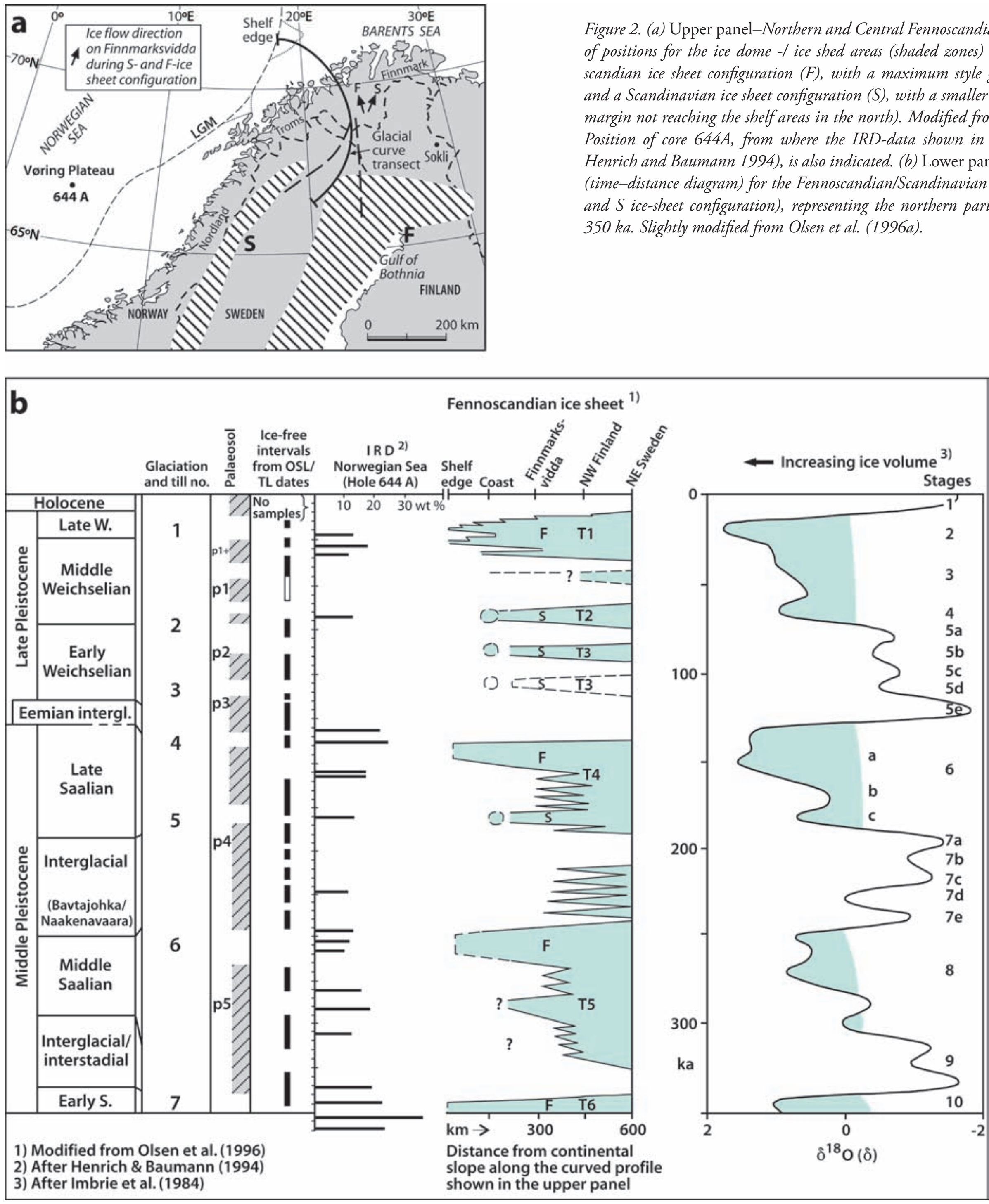


Figure 2. (a) Upper panel–*Northern and Central Fennoscandia with indication of positions for the ice dome -/ ice shed areas (shaded zones) during a Fennoscandian ice sheet configuration (F), with a maximum style glacier extension; and a Scandinavian ice sheet configuration (S), with a smaller ice extension (ice margin not reaching the shelf areas in the north). Modified from Olsen (1988). Position of core 644A, from where the IRD-data shown in (b) derive (after Henrich and Baumann 1994), is also indicated. (b)* Lower panel–*Glacial curve (time–distance diagram) for the Fennoscandian/Scandinavian ice sheet (with F and S ice-sheet configuration), representing the northern part, during the last 350 ka. Slightly modified from Olsen et al. (1996a).*

deep-sea sediments on the Vøring Plateau from 2.74 Ma (Jansen and Sjøholm 1991, Jansen et al. 2000), and is also very close to the onset of the Quaternary (see above). A long record of glacial/interglacial history may be found in the large submarine fans on the continental slope, located at the outlet of troughs where fast-moving ice streams during maximum glaciations ended at the edge of the continental shelf (Figure 3) ('Trough mouth fans', Vorren et al. 1991, King et al. 1996, Laberg and Vorren 1996, Vorren and Laberg 1997, Sejrup et al. 2000). Sediment-core data from the southwestern flank of the North Sea Fan indicate that the ice sheet during the Last Glacial Maximum (LGM) interval terminated at the mouth of the Norwegian Channel in three separate phases between 30 and 19 cal ka (cal = calendar or calibrated) (Nygård et al. 2007).

Based on a dense pattern of seismic lines, and a compilation of previous seismic mapping, Dahlgren et al. (2002) and Rise et al. (2002, 2005a) have recently made a detailed 3D model of the late Cenozoic deposits on the Mid-Norwegian shelf. They

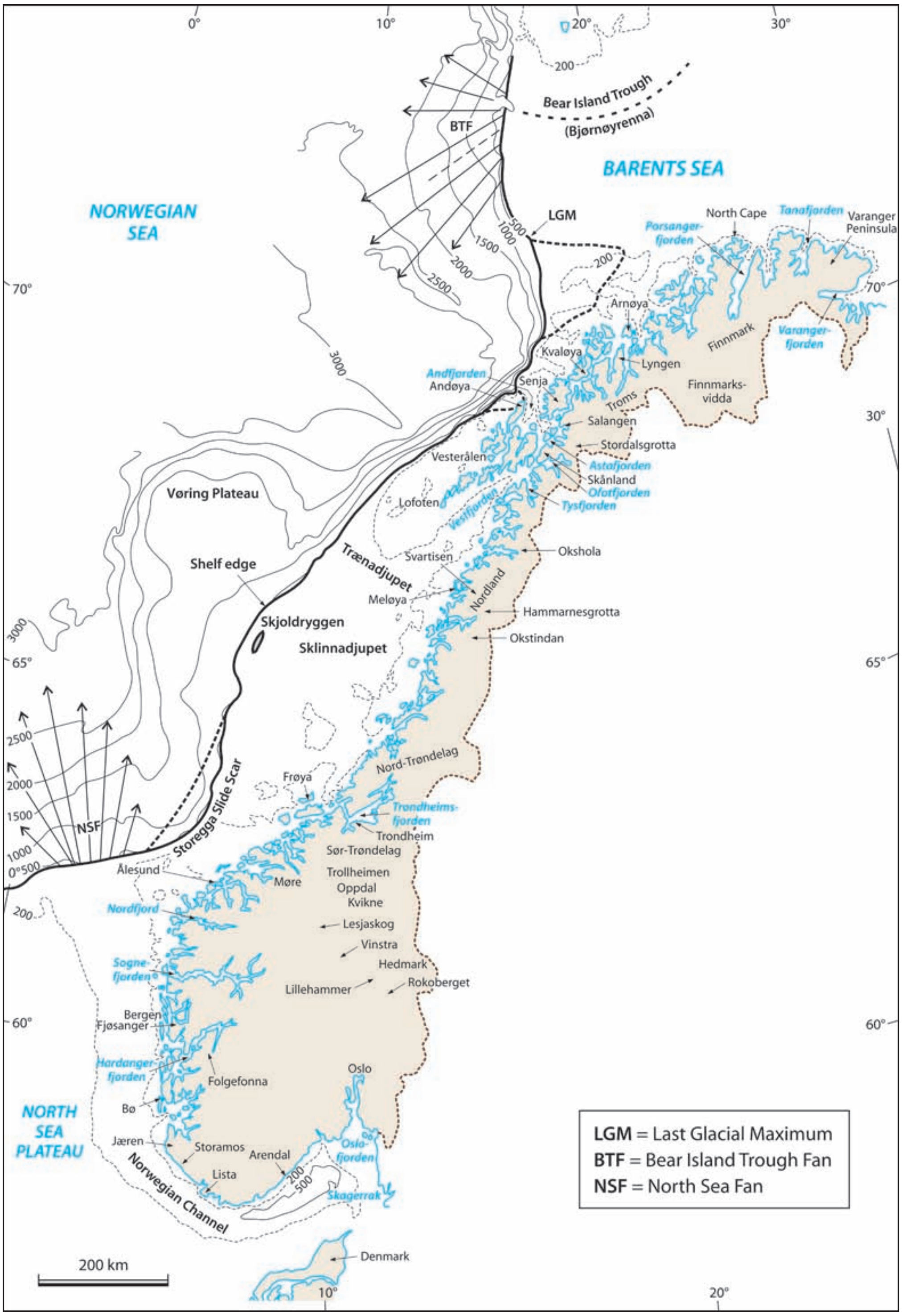


Figure 3. Map of Norway with the continental shelf. The Late Weichselian maximum glacial limit is marked and follows generally the shelf edge. Deviations indicated by stippled lines are discussed in the main text. Geographical names used in the text are indicated, except most of those related to the late-glacial history, which are shown in Figures 22, 23, 24 and 25. Modified from several sources, including Mangerud (2004).

concluded that during all the three last major glaciations (the Elsterian, the Saalian and the Weichselian) the ice sheets reached the edge of the shelf and even extended the shelf westwards with huge accumulations of sediments, and that the Elsterian and the Saalian ice sheets in several areas reached even farther to the west during maximum glaciation than the Weichselian, just as known from Central Europe (e.g., Ehlers 1996). However, in some Mid-Norwegian shelf areas, and in contrast to the Central European record, the Weichselian seems to have been the most extensive of these glaciations to the west (Rise et al. 2005a). During all these glaciations westwards expansion was simply limited by the deep sea beyond the shelf edge. Where the water depth is sufficient, the ice front generally floats, and in some areas possibly making an ice shelf, probably up to at least 200–300 m thick (present shelf-ice thickness of up to 800 m is recorded from the Antarctic) as considered from modern analogues. At least 1/10 of the ice thickness of a grounded ice is above sea level. However, ice walls reaching more than 50–60 m above sea level are not stable and will collapse rapidly (Vorren and Mangerud 2008). Therefore, the ice will float as water depth increases to more than 500–600 m. Further expansion is limited by iceberg calving, which also controls the vertical extent of the ice sheet by increasing the steepness of the ice surface gradient and downdraw of ice masses farther upstream in the mountain areas adjacent to the coastal zone.

The North Sea–Netherlands–Germany–Denmark

The North Sea Plateau and the adjacent Norwegian Channel
Data from the Troll core 8903 (Sejrup et al. 1995) and seismostratigraphical information, also from the Norwegian Channel (Sejrup et al. 2000), indicate at least one Saalian (sensu stricto) and four pre-Saalian major ice advances with deposition of tills. There seems to be a long nonglacial interval with deposition of a thick marine sediment unit between the oldest till bed at *c.* 1.1 Ma and the subsequent till that may have a Marine Isotope Stage (MIS) 12 age (450 ka) (Sejrup et al. 1995, Nygård 2003). Glacial sediments in a core from the Fladen area on the North Sea Plateau, indicate a glacier reaching this area as early as 850 ka (Sejrup et al. 1991), but did not reach a full maximum size at the mouth of the Norwegian Channel at this time (Sejrup et al. 2005).

The Netherlands–Germany–Denmark
Based on the record of the till sheets and end moraines, the maximum and oldest glacier extensions during the last million years in this region are represented by those that occurred during the Elsterian (MIS 12) and Saalian (MIS 6, 140–190 ka) glaciations (Ehlers 1996). Erratic boulders from Scandinavia in even older sediments in these areas may indicate glacier expansions of similar sizes prior to the Elsterian, e.g., as old as the Don glaciation (at least 500 ka, but less than the Matuyama/Brunhes magnetic reversal boundary at *c.* 780 ka) that had a Fennoscandian/Scandinavian ice-sheet origin and is represented by till deposits near the river Don in Eastern Europe (Belarussia, Ehlers 1996).

The mainland of Norway

All significant Fennoscandian ice growths during the Pleistocene were initiated in or close to the Scandinavian mountains. Most field data and several recently published glacial models support this assumption (see review by Fredin 2002). However, the possibility of an *"instantaneous glacierization"* with major ice growth on inland plateaux (Ives 1957, Ives et al. 1975, Andrews and Mahaffy 1976) should not be disregarded, but is difficult to prove since the ice-flow pattern from this type of growth would probably be the same as ice growth from big inland remnants of ice after a partial deglaciation (Olsen 1988, 1989).

Old Quaternary deposits are found in karst caves. About 1100 caves are now known in Norway, many of which contain speleothems (Lauritzen 1984, 1996). Speleothems need ice-free, subaerial conditions to develop (Lauritzen et al. 1990, Lauritzen 1995, Linge 2001). Therefore, such deposits constrain the reconstruction of glaciers, since these cannot have extended over the hosting caves during speleothem growth. The oldest well-dated speleothem yielded an age of 500 ka (MIS 13) (U-series dating with mass spectrometry), and even older, possibly pre-Quaternary caves exist (Lauritzen 1990). However, most U-series dating of speleothems have given Eemian or younger ages (Lauritzen 1991, 1995).

Ice-damming conditions with deposition of fine-grained, laminated sediments in coastal caves have been demonstrated to be an important indicator for glacier expansion, which has occurred and reached beyond the cave sites at least four times during the Weichselian, whereas occurrences of blocks falling from the roof of the caves during ice-free conditions clearly constrain the corresponding glacier expansions (Larsen et al. 1987) (Figure 4).

Pre-Saalian glacial and interglacial deposits are so far found in two areas: 1) Finnmarksvidda in North Norway (Figure 3) is a rolling (wavy, low-relief) plain of moderate altitude *c.* 300–350 m a.s.l., with thick Quaternary deposits (maximum thickness >50 m, and average thickness estimated to *c.* 6 m) including till beds,

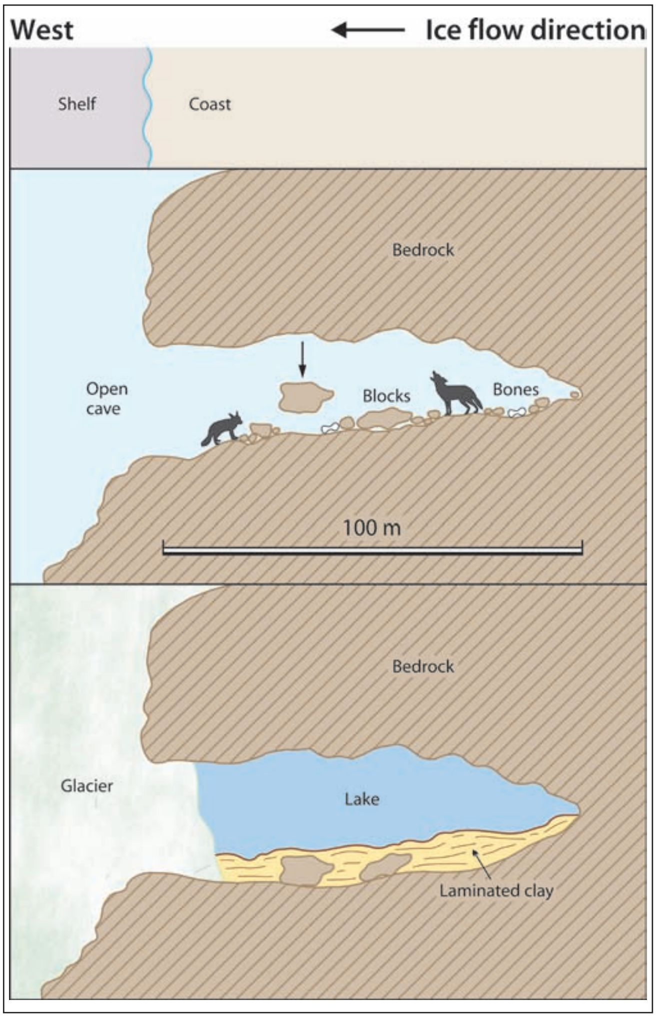
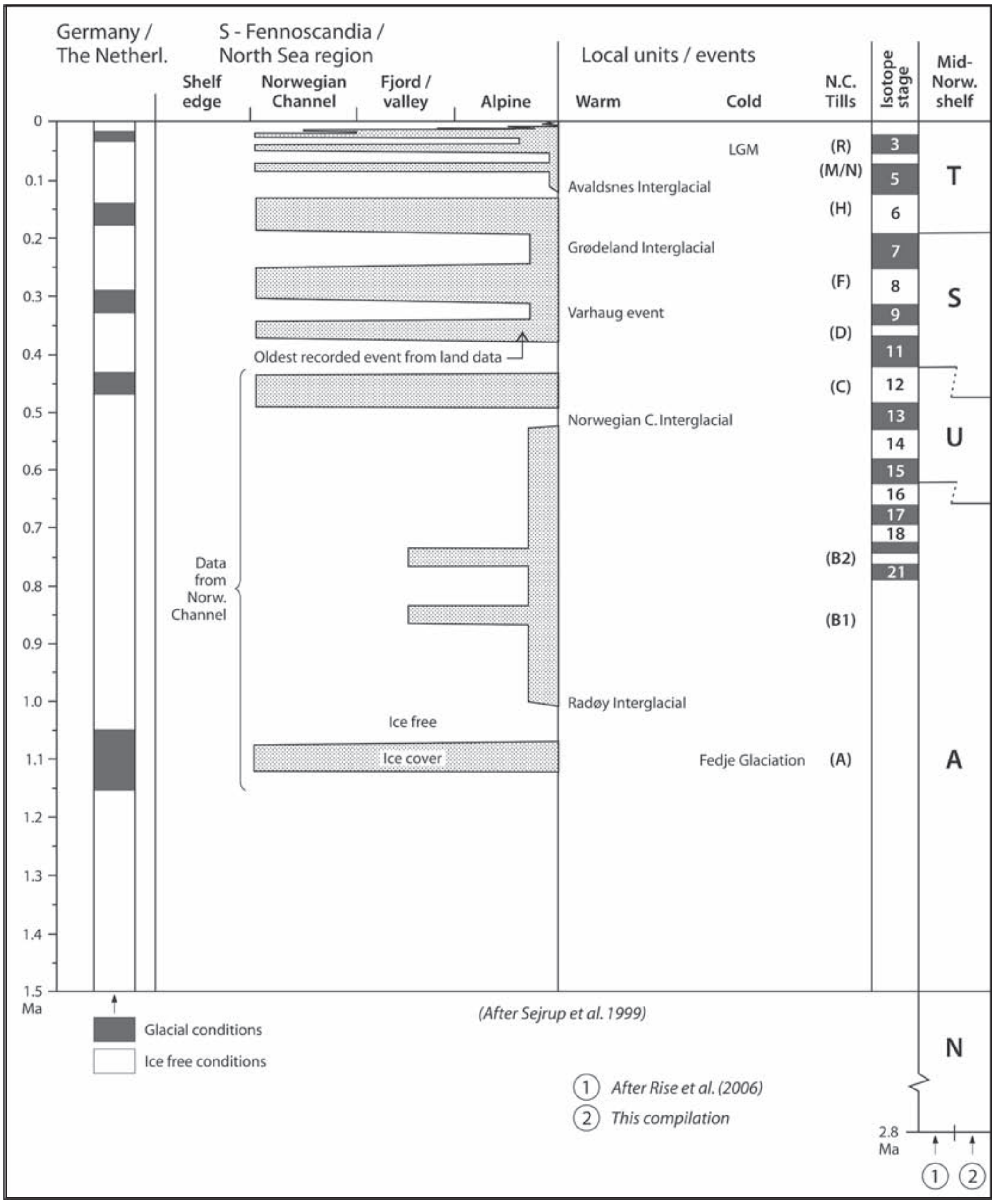


Figure 4 Conceptual cross sections of uplifted, marine caves with sediments in western Norway. The upper and lower sketches represent ice-free and ice-dammed conditions, respectively. Modified from Mangerud et al. (2003).

Figure 5. Glaciation curve for the Fennoscandian ice sheet, representing the southern part, including the North Sea region, during the last 1.1 million years. The stratigraphical positions of the till units A, B1, B2, C, D, F, H, M/N? (uncertain stratigraphical position) and R from the Norwegian Channel (N.C.), and the correlations with the major glacial events in Germany and The Netherlands and with the marine isotope stages are also indicated. The oldest recorded event from the mainland is represented by the till that is correlated with Norwegian Channel unit D (Isotope stage 10). Modified from Sejrup et al. (2000, 2005). Correlations with the Mid-Norwegian shelf area are shown to the right (after Rise et al. 2006, and this compilation).



interstadial and interglacial deposits, and soils (palaeosols) dating back to at least MIS 9 or 11 (Figure 2b) (Olsen 1998, Olsen et al. 1996a, Larsen et al. 2005). 2) Even thicker Quaternary deposits occur in the Jæren lowlands in southwest Norway. Here glacial, interstadial and interglacial deposits with marine fossils, from MIS 10 and upwards are described, partly from sections, but mainly from boreholes (Figure 5) (Sejrup et al. 2000). Elsewhere in Norway no glacial deposits are proven older than the Saalian (MIS 6), but till of that age has been found in the inland (Olsen 1985, 1998), and even (in sheltered positions) in the fjord areas (below Eemian sediments), where the glacial erosion generally has been most intense (Mangerud et al. 1981b, Aarseth 1990, 1995).

Glaciation curve and glacier extension maps

The glaciation curve for northern Fennoscandia (Figure 2b) is used here as an illustration of glacier variations for the Fennoscandian/Scandinavian ice sheet. The curve is drawn as a time–distance diagram along a transect from central northern Sweden, across northern Finland and northern Norway (Finnmark) and further on to the shelf area in the northwest. The transect follows the major ice-flow directions during moderate to large ice-volume intervals. The coloured zones in the diagram indicate the ice-covered areas (horizontal axis) at a particular time (vertical axis) during the last 350,000 years.

Stratigraphical data mentioned above are in this illustration

(Figure 2b) combined with IRD data from the Norwegian Sea (Henrich and Baumann 1994) and average ice-volume data from the North Atlantic region (Imbrie et al. 1984). Only the highest IRD production values (>10 wt.%) are included, so that the small IRD peaks, for example during those parts of the last glaciation (Weichselian) which have minor ice volume, are not shown. Nevertheless, there is a clear correspondence in medium to high IRD compared with medium to high glacier extensions/ volumes. An exception to this trend might be during MIS 8, where high IRD around 280 ka apparently corresponds to a moderate ice extension and volume, whereas moderate IRD at *c.* 260 ka apparently corresponds to a major ice extension and volume. However, a combination of all these results indicates that maximum glacier extensions, with the ice-sheet margin at the shelf edge in the northwest, probably occurred at least three times between the Holsteinian (MIS 11) and Eemian (MIS 5e) interglacials.

No tills from the pre-Holsteinian period are confirmed to occur in northern or central Fennoscandia so far.

Glacier extension maps show the maximum glacier extensions areally, with the assumed position of the ice margin indicated. Illustrations of such maps are included for three pre-Weichselian stages in Appendix A (Figure A1), and also included for several Weichselian stages, which we will deal with later (see below).

The Eemian

The last interglacial, the Eemian in the northern European Quaternary terminology, was generally warmer than the present interglacial (Holocene), and it is represented by marine and/ or terrestrial organic and/or other deposits/formations recorded at several sites on the mainland (Vorren and Roaldset 1977, Sindre 1979, Mangerud et al. 1981, Andersen et al. 1983, Aarseth 1990, Lauritzen 1991, Olsen et al. 1996a, Olsen 1998), as well as on the shelf (e.g., Haflidason et al. 2003, Hjelstuen et al. 2004, Rise et al. 2005a, b). The Eemian is one of the most important stratigraphical markers in the Late Pleistocene history, and particularly since it was suggested (and later shown) to correspond with MIS 5e (e.g., Mangerud et al. 1979) it has been the backbone for the stratigraphical framework and most Middle to Late Pleistocene correlations in Norway, including the adjacent shelf. A distribution of all Eemian sites on land and some offshore sites is shown in Figure 6. For more details, see Olsen et al. (2013).

The Early and Middle Weichselian

The glacial/interstadial history and dating problems

The Late Weichselian ice sheet removed most of the older deposits from the mainland of Norway and redeposited it fragmentary or as integrated components in younger deposits around the border zones in the north, west and south. Therefore, only small parts of the pre-Late Weichselian history in Norway are known. The interpretation of the older part of the Weichselian is consequently based on observations from very few localities, so the reconstruction of ice-sheet limits at different stages during the Early and Middle Weichselian is obviously only tentative.

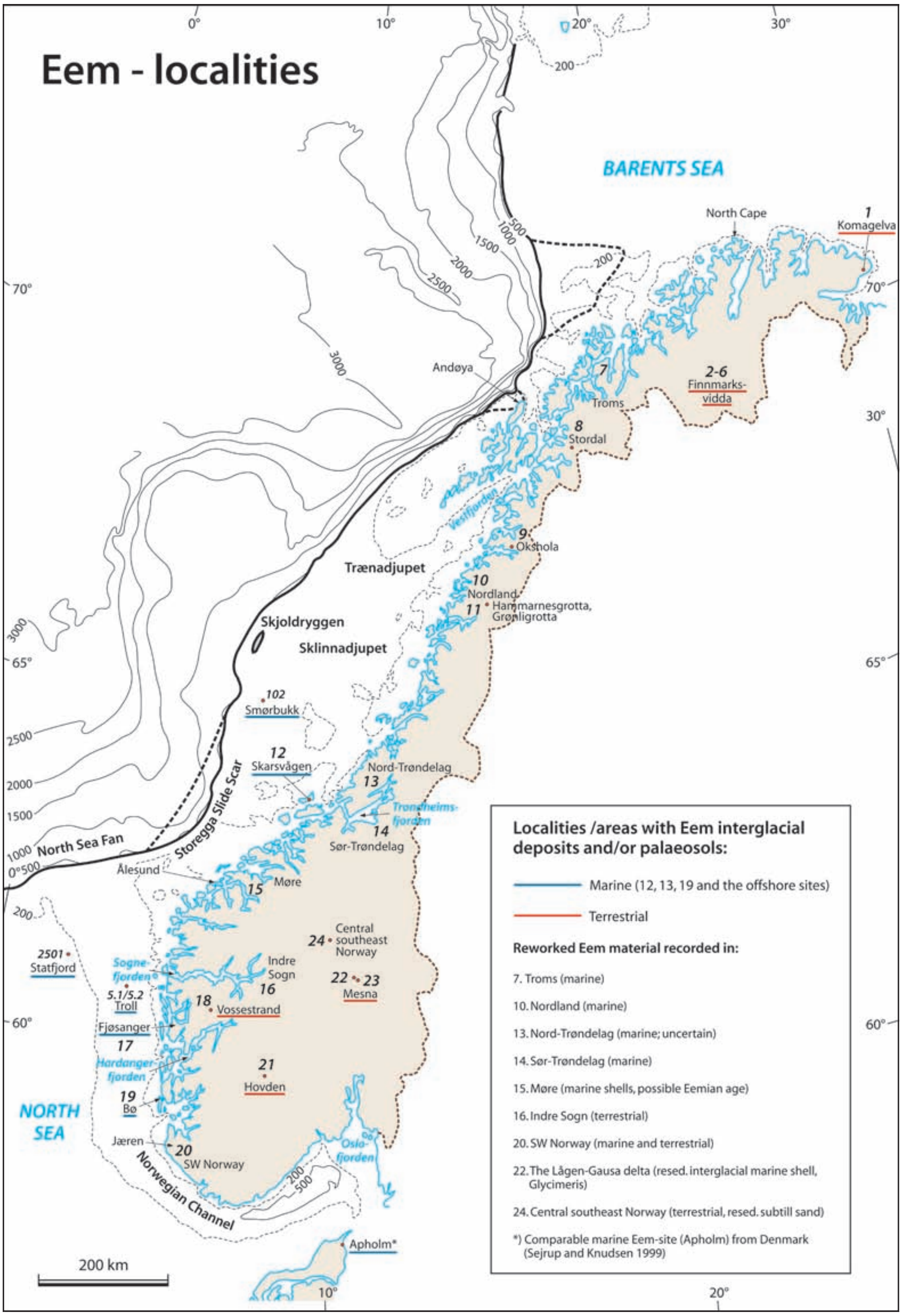
Another uncertainty is the correlation of events based on absolute dating methods. Dates older than the range of the radiocarbon method (max. 40,000–45,000 yr) are particularly problematic. Therefore, the age of the deposits, the correlation between different sites, and consequently also the conclusions on the glacial history, have been controversial, both between different scientists and between steps of research based on different dating methods. Examples of such controversies may be the development of glaciation curves for southwest Norway (Mangerud 1981, 1991a, b, Larsen and Sejrup 1990, Sejrup et al. 2000, Mangerud 2004) and North Norway (Olsen 1988, 1993a, b, 2006, Olsen et al. 1996a). The (essentially presented) two curves from southwest Norway (Figure 7 and Appendix B, Figure B1) are based on the same continental data, and there is a general agreement between these curves in the younger part, i.e., during the post-Middle Weichselian. However, there are considerable differences in the older part of the curves, where Figure 7 (Mangerud 2004, Mangerud et al. 2011) is more tuned to IRD data from deep-sea sediments (Baumann et al. 1995), and also supported by recently reported IRD data (Brendryen et al. 2010), whereas the other curve (Appendix B, Figure B1) (Larsen and Sejrup 1990, Sejrup et al. 2000) relies more directly on amino-acid racemisation (AAR) analyses. Mangerud (2004) explained these changes by a more regional climatically based interpretation (ice-free conditions, marine and terrestrial biological characteristics) of his curve (Mangerud et al. 1981b, Mangerud 1991a, b).

The glaciation curve from North Norway (Figure 8) has been changed several times since it was originally introduced (Olsen 1988), in accordance with input of new dates and knowledge from Finnmark, and also added information from neighbouring areas, e.g., speleothem dates from caves in Nordland and Troms (Lauritzen 1991, 1995, 1996), and sedimentary stratigraphy and dates from Sokli in Finland (Helmens et al. 2000, 2007). The revised curve indicates that the ice sheet did not reach the northern coastal areas during the initial part of the Weichselian. It may have reached these areas at *c.* 90 ka, but more likely *c.* 60 ka and *c.* 44 ka.

Early/Middle Weichselian glacial limits and interstadials/ice-retreat intervals

It is clear from the above examples that there is so far no consensus on the Early and Middle Weicshelian glacial and interstadial history of Norway. This is due to scarcity of known sites of this age (e.g., Mangerud 2004). The first reconstruction of the westward extension of the Fennoscandian ice sheet during the

Figure 6. All reported Eemian sites on land in Norway and selected Eemian offshore sites. Onshore sites: 1–6 – Olsen et al. (1996a), Olsen (1998); 7 – Vorren et al. (1981); 8–9 – Lauritzen (1995); 10 – Olsen et al. (2001a), Olsen (2002); 11 – Linge et al. (2001); 12 – Aarseth (1990, 1995); 13–14 – Olsen et al. (2001a, 2002); 15 (Eidsvik, Møre) – Miller and Mangerud (1985); 16 – Vorren (1972); 17 – Mangerud et al. (1981b); 18 – Sindre (1979); 19 – Andersen et al. (1983), Sejrup (1987); 20 – reviewed by Vorren and Mangerud (2008); 21 – Vorren and Roaldset (1977); 22 – Olsen and Grøsfjeld (1999); 23 – Olsen (1985b, 1998); 24 – Myklebust (1991), O.F. Bergersen (pers. comm. 1991). Offshore sites: 102 (Smørbukk) – Rokoengen (1996); 2501 (Statfjord) – Feyling-Hanssen (1981); 5.1/5.2 (Troll) – Sejrup et al. (1989); other data – Haflidason et al. (2003), Hjelstuen et al. (2004), and Rise et al. (2005a, b).



Early (MIS 5d and 5b) and Middle (MIS 4–3) Weichselian was based on only four sites/areas (Figure 3, Jæren, Bø, Fjøsanger and Ålesund) (Andersen and Mangerud 1989). Very few well-dated sites of this age have been recorded in Norway since then. Therefore, today areal reconstructions of maximum extension of ice sheets of this age in southern Norway still have to be based mainly on information from these few coastal sites/areas, and from additional new data from the continental shelf (e.g., Nygård 2003, Rise et al. 2006, Nygård et al. 2007), a few inland sites in southeast Norway (Vorren 1979, Helle et al. 1981, Olsen 1985b, 1998, Haldorsen et al. 1992, Rokoengen et al. 1993a, Olsen et al. 2001a, b, c) and on general considerations.

The combination of previous and new information from Jæren shows that in southern Norway the ice sheet during MIS 5–3 twice grew to a size allowing development of an ice stream in the Norwegian Channel (Larsen et al. 2000). Primary

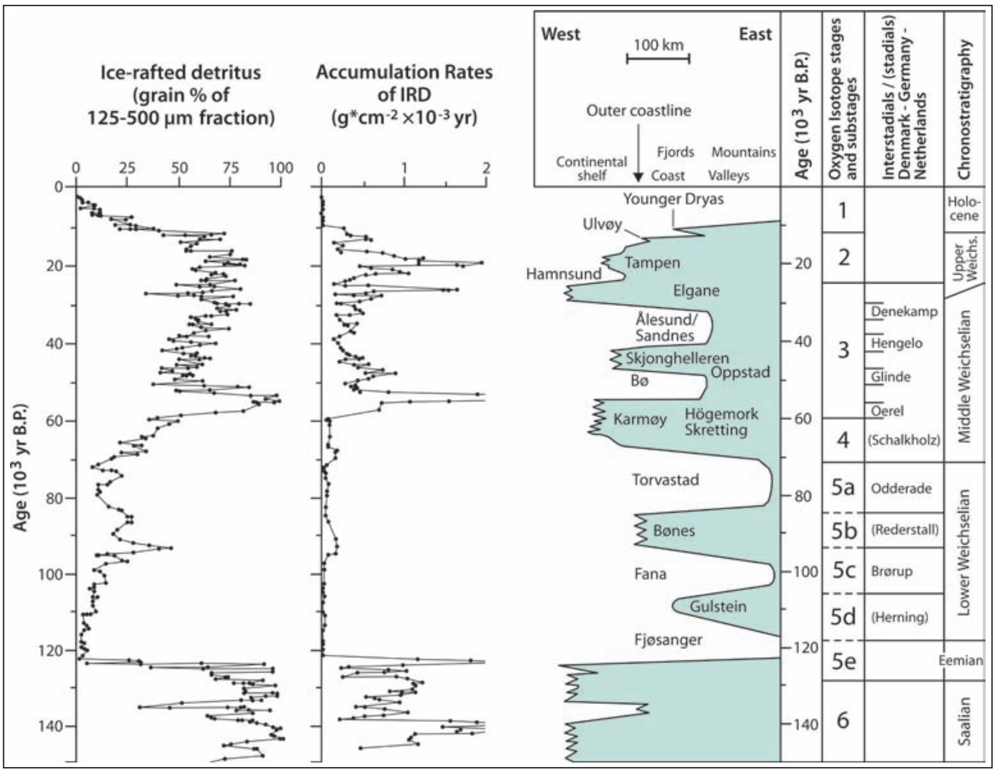


Figure 7. Comparison between IRD data from the Vøring Plateau west of the Norwegian coast and a glacial curve for southwest Scandinavia, west of the watershed and ice-divide zone. The time scale is in calendar years. Modified after Mangerud (2004) and Baumann et al. (1995).

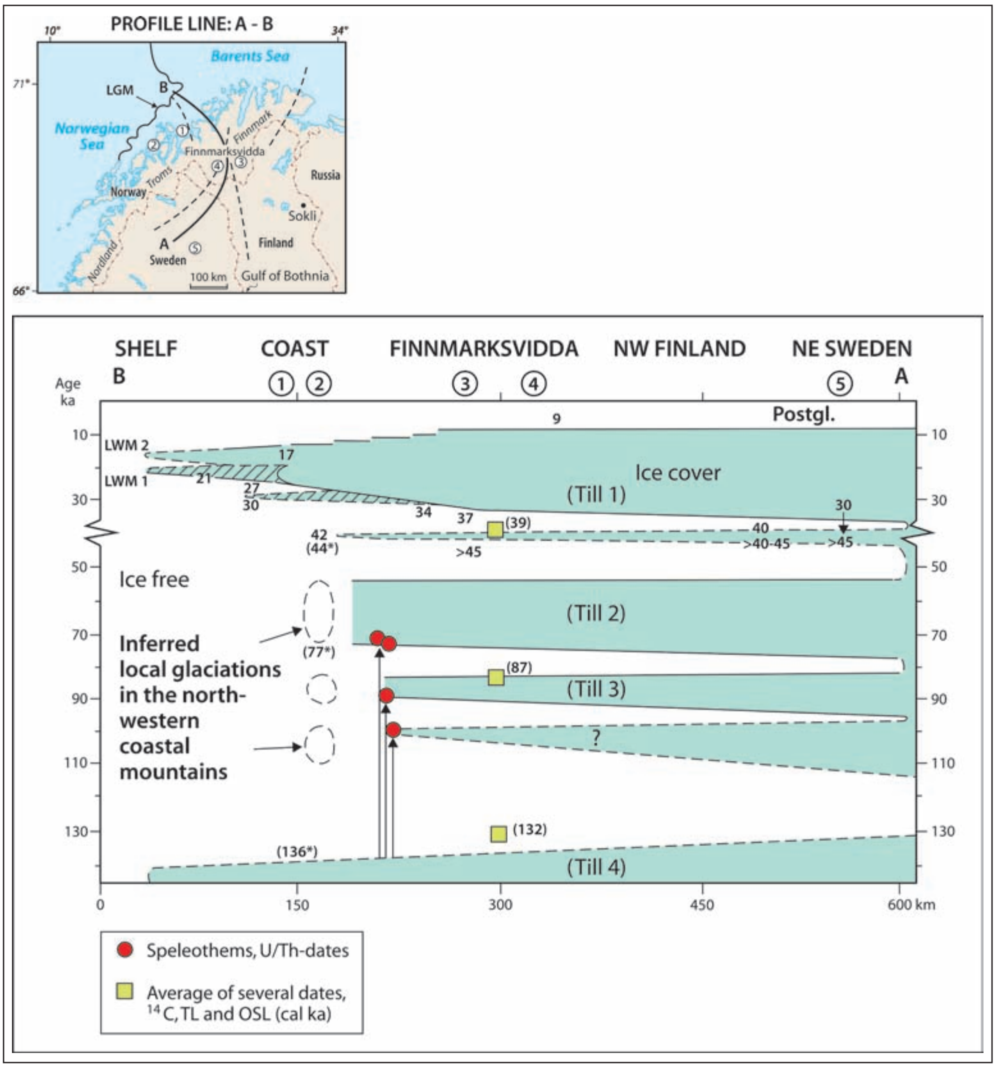


Figure 8. Glacial curve for the last 145,000 yr in northern Fennoscandia. Upper panel: *Location of profile line A–B and the stratigraphical site areas 1–5. The Last Glacial/Late Weichselian Maximum (LGM/LWM) limit at the shelf edge is indicated in the NW.* Lower panel: *Glacial curve for the last 145,000 yr in northern Fennoscandia, with dates (^{14}C, AAR, TL and OSL) in cal ka (in parantheses) and in ^{14}C ka. Note the change in time scale at c. 40 ka (^{14}C yr up to this age and calendar yr for older ages). Speleothem data is from Troms and Nordland, and are briefly mentioned in the main text. *Dates from redeposited material. Updated version after Olsen (2006), modified from Olsen et al. (1996a) and Olsen (1988).*

and extensive descriptions of the terrestrial coastal sites in the southwest exist for Jæren (Andersen et al. 1987, Larsen et al. 2000, Sejrup et al. 1998, Raunholm et al. 2004), Bø on Karmøy (Ringen 1964, Andersen et al. 1983, Sejrup 1987), Fjøsanger (Mangerud et al. 1981b), and the Ålesund area (Larsen et al. 1987, Mangerud et al. 1981a, 2003, Valen et al. 1996).

A new reconstruction of the extension of the ice sheet at different intervals of the Weicshelian for Norway and adjacent areas is presented by Olsen (2006) (Figure 9a). The ice-sheet variations in the Barents Sea are not indicated, but during maximum ice extension the Fennoscandian ice sheet is supposed to have moved independently, but partly coalesced with

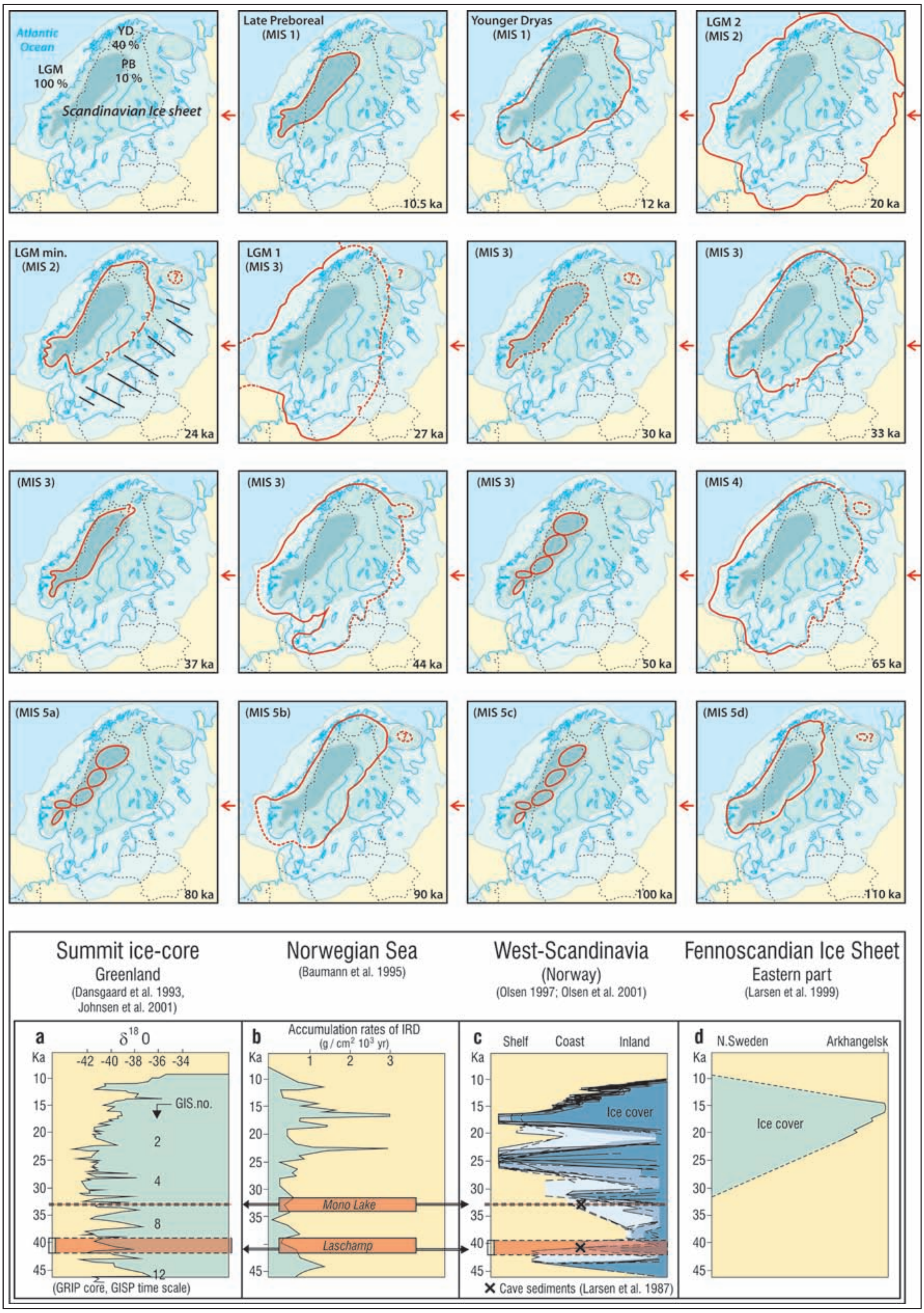


Figure 9. (a) Upper panel – Extension of the Scandinavian ice sheet during different stages of the Weichselian glaciation. The 16 maps are based on data from Lundqvist (1992), Larsen et al. (1999), Lunkka et al. (2001), Olsen et al. (2001a, b, c), Mangerud (2004) and Olsen (2006). The LGM ice extension is marked on all maps without the coalescing zones with ice sheets in the southwest (British Isles) and the northeast (Barents Sea – Kara Sea). Key map, with ice extension in area% of LGM for Younger Dryas and Preboreal, is included to the left in the uppermost row. Shading with parallel lines during LGM minimum (Trofors interstadial) indicates a possible, more or less continuous ice advance in the eastern sector between LGM 1 and LGM 2. Ages in ka, here used as cal ka. (b) Lower panel – Comparison between various glacial records, including correlations based on palaeomagnetic signals (Mono Lake, Laschamp). Diagram c is based on glaciation curves from Norway along nine transects showing regional trends from inland to shelf. Ages refer to cal ka.

the Barents Sea ice sheet in the southwesterns Barents Sea area (Landvik et al. 1998, Vorren and Mangerud 2008). This reconstruction is a modified version of the reconstructions by Lundqvist (1992) and Mangerud (2004), and includes both stadials and interstadials (ice-retreat intervals, without vegetational data). The reconstruction is based on e.g., the evaluation by Mangerud (2004), concluding that the ice front passed beyond the coast near Bergen (Fjøsanger) during MIS 5b. Supposedly it crossed the coast over a wider area during MIS 4 and 3, since an ice stream may have developed in the Norwegian Channel during these events, demonstrating that the ice limit was well outside the coast of southern Norway at that time (Larsen et al. 2000). In Mid-Denmark, several occurrences of till with Scandinavian erratics represent the Sundsøre glacial advance from Norway, which has been dated to 60–65 ka (Larsen et al. 2009a, b) and, therefore, suggest a considerable MIS 4 ice extension well beyond the coastline of Norway in the south and southwest. The subsequent Ristinge ice advance from the Baltic Sea is dated to *c.* 44–50 ka, i.e., early MIS 3, and is also represented by many occurrences of till across much of Denmark (Houmark-Nielsen 2010), which implies that the ice extension must have reached far beyond the coast of southern Norway also during this time. In a compilation of the glacial variations in southwest Norway, the maximum glacier expansions during MIS 4 and 5 are restricted to the fjord region and without glacial debris-flow (GDF) formation at the North Sea Fan (Sejrup et al. 2005). However, recent studies from the North Sea Fan indicate that the Weichselian glacier expansion possibly reached to the mouth of the Norwegian Channel and triggered a glacial debris flow there once before the LGM, and that may have occurred as early as during MIS 5b (Figure 5) (Nygård et al. 2007, fig.2, GDF P1d cut reflector R2 which is estimated to 90 ka). A problematic unit, a till (M/N), is located in the North Sea Fan and deposited by an ice stream in the Norwegian Channel. It was originally favoured by Sejrup et al. (2000) to be from the Early Weichselian, and re-evaluated to a MIS 4 age by Mangerud (2004), but is considered here as undated and not assigned a particular age.

At Skarsvågen on the Frøya island (Figure 3), Sør-Trøndelag, Eemian terrestrial sediments (gyttja) are overlain by Early Weichselian marine transgression sediments followed by regression sediments and till, which is covered by a Middle to Late Weichselian till on top (Aarseth 1990, 1995). The marine sediments overlying the Eemian gyttja there derive from a deglaciation, which may succeed a glaciation from MIS 5d, 5b or 4, perhaps with the latter as the most likely alternative, since a proper MIS 5 till, as far as we know, has not yet been reported from the coastal part of Central Norway. In that case the overlying till is likely to be of MIS 3 (44 ka) age, whereas the younger till bed may be either of MIS 3 (34 ka) or MIS 2 age. However, the age problem for these deposits is not yet solved (I. Aarseth, pers. comm. 2004).

At Slettaelva on Kvaløya (Figure 3), northern Troms, the sediment succession starting at the base on bedrock includes a till from a local glacier trending eastwards, i.e., in the opposite direction compared to the subsequent Fennoscandian ice sheet (Vorren et al. 1981). The till is overlain by ice-dammed sediments caused by the advance of the Fennoscandian ice sheet during a period sometime before 46 ka (Vorren et al. 1981), possibly during MIS 4 *c.* 60–70 ka (Figure 8). On top of these sediments follows a till which is divided in three subunits that each may represent an ice advance over the area. The lowermost of these contains shell fragments, which are dated to 46–48 ka. Consequently, three glacier advances of which the oldest may be *c.* 44 ka have reached beyond this site, and one earlier, possibly local ice advance may have reached to the site.

Continuous speleothem growths from the last interglacial to 100, 73 and 71 ka have been reported from different caves from the inner fjord region of northern Norway (Figure 3; Stordalsgrotta, Troms, 920 m a.s.l., Lauritzen 1995; Hammarnesgrotta, Rana, 220 m a.s.l., Linge et al. 2001; Okshola, Fauske, 160 m a.s.l., Lauritzen 1995). This gives clear constraints for the westward (and vertical) extension of glaciations both during MIS 5d and 5b (Figures 8 and 9a), because such growth needs subaerial, humid and nonfrozen conditions and cannot proceed subglacially or under water (Lauritzen 1995).

At Leirhola on Arnøya (Figure 3), northern Troms, till-covered *glaciomarine* deposits up to *c.* 8 m a.s.l. indicate that the glacier margin was close to the site a short time before *c.* 34–41 ka (i.e., probably around 40 ^{14}C ka), advanced beyond the site with deposition of a till shortly after *c.* 34 ka, and retreated and readvanced over the site with additional till deposition shortly after 31 ka (Andreassen et al. 1985). This is only one of many similar sites on islands and on the mainland along the coast of North and Central Norway where ^{14}C-dates of marine shells in tills or subtill sediments indicate an ice advance *c.* 44 ka that reached beyond the coastline (e.g., Olsen et al. 2001c, Olsen 2002). The best record of an ice advance of this size and age is from Skjonghelleren near Ålesund (Figure 3). A correlation with the Laschamp geomagnetic excursion suggests quite strongly that an ice advance reached beyond this coastal cave *c.* 44 ka (Larsen et al. 1987).

Indirect indications of earlier large ice volumes, for example data which have the implication that a considerable depression still was present due to glacial isostasy, are represented at some elevated sites in southeast Norway, e.g., at Rokoberget (Rokoengen et al. 1993a, Olsen and Grøsfjeld 1999) where a major ice advance prior to 38 ka, probably at *c.* 44 ka, is inferred based on dated highly raised sediments with marine microfossils.

A considerable MIS 3 ice retreat and deglaciation before and after the *c.* 44 ka ice advance is assumed based both directly and indirectly on field data. For example, reported interstadial sites from the Bø interstadial, the Austnes interstadial and the Ålesund interstadial (e.g., Mangerud et al. 2010), represent direct indications of reduced ice extension, whereas e.g., low MIS 3 shorelines, which have been reported from North Norway (Olsen 2002, 2010), indicate indirectly a minor MIS 3 ice volume and thereby also a minor ice extension.

The vertical extent of the last major ice sheet is discussed by e.g., Nesje et al. (1988) and Brook et al. (1996). They used trimlines (boundary between autochthonous blockfields and ground covered by glacial deposits) and dates from cosmogenic nuclide surface-exposure dating to constrain the vertical LGM ice extent. However, neither of these methods excludes the possibility of an LGM ice cover of cold-based ice, but they can both be used to exclude significant erosion on the highest mountains during the LGM interval. The vertical ice extension is dealt with below.

A composite curve based on a combination of curves along nine transects from inland to coast and shelf from different parts of Norway demonstrates clearly the regional synchroneity of the Middle and Late Weichselian ice-sheet variations of the western part of the Scandinavian ice sheet (Figure 9b, diagram c). A comparison with other proxy glacial and climatic data in the vicinity of Norway, also shown in Figure 9b (lower panel), strengthens the character of regional synchroneity of glacial and climatic reponses, and indicates that these responses reach much further than within the boundaries of Norway.

Olsen (2006) recently presented a reconstruction of relative size variations, both in area and ice volume, for the Scandinavian ice sheet during the Weichselian glaciation (Appendix B, Figure B2). As expected, the ice growth in horizontal and vertical direction seems to match fairly well in some intervals, but not in all. For more details, see Appendix B.

The Late Weichselian Glacial Maximum (LGM)

The ice-sheet limit on the Norwegian Shelf

An extensive review and synthesis of the Quaternary geology on the Norwegian Shelf was given by H. Holtedahl (1993). The information since then is compiled most recently by D. Ottesen in his doctoral thesis (Ottesen 2007). General conclusions and many of the references to literature used here refer to these publications.

The LGM interval is here subdivided in three phases, LGM 1 (>25–26 ka), LGM minimum, corresponding to the Andøya–Trofors interstadial (Vorren et al. 1988, Olsen 1997), and LGM 2 (<20–21 ka). Further details of LGM 1 and 2 are given later under the heading *The age of the glacial limits*.

At present there is consensus among geologists working on the Norwegian Shelf that the Late Weichselian ice sheets during maximum extension reached the shelf edge along the entire length from the Norwegian Channel to Svalbard (e.g., Rokoengen et al. 1977, Andersen 1981, Holtedahl 1993, Mangerud 2004, Ottesen et al. 2005a). The only suggested exception that still seems to gain some support is a narrow zone west of Andøya (Figure 3), where the limit of the active coherent ice sheet, according to field data and dates in both previously (Vorren et al. 1988, Møller et al. 1992) and the most recent compilation (Vorren and Plassen 2002), was located on land (on the island).

The glacial limit is mapped using the same three criteria as listed by Mangerud (2004): 1) by the western limit of the youngest till sheets (for most of their length), or the inferred 'grounding line' of till tongues, mainly recorded by seismic methods, 2) end-moraine ridges, and 3) submarine fans built up mainly by debris flows of glacial sediments.

A long-standing debate on whether the Scandinavian and British ice sheets met in the North Sea during the LGM interval and where the ice margins were located advanced significantly with the data presented by Sejrup et al. (1994, 2000) and Carr et al. (2000, 2006). They concluded that the ice sheets probably met and that these maximum positions were reached between 33 and 26 ka. Moreover, based on seismic signatures at the mouth of the Norwegian Channel, King et al. (1996, 1998) placed the LGM limit in this area at the boundary between basal till in the channel and the glacial debris flows on the North Sea Fan, and they dated this limit to 27–19 ka (King et al. 1998). Recently, Nygård et al. (2007) found from similar data that the ice sheet reached the maximum position in this area three times during the LGM interval (30–19 ka).

The LGM limit on the Mid-Norwegian shelf follows the shelf edge. The outermost till has partly been eroded by the huge Storegga Slide (Bugge et al. 1978, Bugge 1980, Jansen et al. 1987, Haflidason et al. 2005, Hjelstuen et al. 2005), and the LGM limit may have been up to 60–70 km west of the present shelf edge (stippled line in Figure 3).

The LGM position in the southwestern corner of the Barents Sea is also a matter of debate. Vorren and Kristoffersen (1986) mapped end moraines on the shelf some 100–150 km from the shelf edge off Tromsø and Finnmark counties (stippled line in Figure 3) and suggested these to represent the LGM limit there. However, recent compilations consider the LGM limit also in this area to be at the shelf edge during coexistence and confluence in the Bjørnøya Trench with the Barents Sea ice sheet (e.g., Landvik et al. 1998, Svendsen et al. 2004).

The age of the glacial limits

Radiocarbon dates on bones from the Skjonghelleren cave, which is situated with its bedrock floor at 30–45 m a.s.l. (i.e., approximately up to the late-/postglacial marine limit), including more than 30 ^{14}C–AMS dates in the range 33–39 cal ka, indicate a maximum age for a late Middle Weichselian ice advance that crossed the coast at Møre after the Ålesund interstadial (Figures 3 and 7, Mangerud et al. 1981a, Larsen et al. 1987, Baumann et al. 1995). This advance is also dated by the record of the 32–33 ka Lake Mungo (Australia)/Mono Lake (British Columbia) palaeomagnetic excursion in clay deposited in a lake dammed in the cave by this ice advance (Larsen et al. 1987). Therefore, an ice front passed the coast near this cave at *c.* 33 ka. These results were supported by similar data from a nearby cave, Hamnsundhelleren (also above the late-/postglacial marine limit) (Valen et al. 1996). However, this was not necessarily a major LGM ice advance, because from the latter cave, two dates of *c.* 28.5 ka were also obtained on animal bones under ice-dammed sediments, indicating a withdrawal of the ice front followed by a significant readvance that, this time, probably extended to the maximum position at the shelf edge. The first and largest LGM ice advance from Norway to Denmark in the south has been dated to *c.* 27–29 ka (Houmark-Nielsen and Kjær 2003, Larsen et al. 2009a), which seems to correspond with a first major (LGM 1) ice advance after the Hamnsund interstadial in western Norway, that terminated on the shelf edge possibly a thousand years or more later, i.e., about 26–28 ka.

Dates of various materials that can be used to bracket the age of the glacial maximum are reported from the areas near the coast, and by stratigraphical correlation also from the fjord valleys and inland areas of most parts of Norway (Olsen 1997, Olsen et al. 2001a, b, c, d, 2002). They constrain *the culmination* of an initial LGM advance to *c.* 34–33 ka, the first major LGM advance (i.e., LGM 1) to *c.* 27–25 ka, and a major readvance to *c.* 20–18.5 ka (i.e., LGM 2). On the western flank of the Scandinavian ice sheet such dates are scarce, but some exist. For example, a set from the North Sea (Appendix B, Figure B3) indicates radiocarbon dated ages between 33–27 ka for the maximum (LGM 1), and between 22.5–18.6 ka for a major readvance (LGM 2) (Sejrup et al. 1994, 2000) that produced an ice stream in the Norwegian Channel. This ice stream terminated for the last time in maximum position at the mouth of the channel and triggered deposition of debris flows on the North Sea Fan between 20 and 19 ka (Nygård et al. 2007).

Another, but smaller readvance (Late Weichselian Karmøy readvance) between 18.6 and 16.7 ka has been recorded, with drumlinised marine and glacial deposits at Bø on Karmøy, southwest Norway (Figures 3 and 10) (Olsen and Bergstrøm 2007). Similar dates and ice-margin fluctuations have been reported from studies of lake sediments on Andøya in northern Norway (Vorren 1978, Vorren et al. 1988, Alm 1993). The data from these lakes have been correlated with those in the adjacent fjord and on the shelf (Figure 11) (Vorren and Plassen 2002), where LGM 1 is represented by Egga I. This is supported by a high maximum sea level that followed Egga I (Vorren et al. 1988), new morphological data from the shelf area (Ottesen et al. 2005), and the LGM 2 readvance, which in this area also reached to or almost to a maximum position during the local Egga II-phase. This latter part of the reconstruction is supported by exposure dates, mainly from bedrock surfaces from Andøya and adjacent areas, by Nesje et al. (2007). The Bjerka readvance is supposed to have occurred shortly before Egga II and did not reach fully to the shelf edge. A subsequent smaller readvance after Egga II is named Flesen.

During the LGM interval the ice margin seems to have reached its maximum (or almost maximum) westerly position twice in the northwest and west (Figures 3 and 11) (Nordland and Troms) (Vorren et al. 1988, Møller et al. 1992, Alm 1993, Olsen et al. 2002), but only once (LGM 1) in the southwest, where LGM 2 (Tampen readvance) was significantly less extensive, but still big enough to produce an ice stream in the

Figure 10. Map with ice-flow indicators (drumlins, striations) and the location in a westward-trending drumlin of the stratigraphical site Bø, Karmøy island, southwest Norway. Modified from Ringen (1964). Stratigraphical data from the Bø site refer to Andersen et al. (1983), Sejrup (1987), and Olsen and Bergstrøm (2007).

Norwegian Channel (Sejrup et al. 1994, 2000). The readvance (LGM 2) in western Scandinavia seems to correspond well with the maximum extension in the east (c. 17 cal ka), in northwest Russia, demonstrating the dynamic behaviour of the ice sheet and the time-transgressive character of the ice flows (Larsen et al. 1999, Demidov 2006).

Lehman et al. (1991), King et al. (1998) and Nygård et al. (2007) reported dates of glacial debris flows deposited from the ice front onto the North Sea Fan (Figure 12). They found several debris flows from the LGM interval, and according to Lehman et al. (1991) a thin debris flow unit on top indicates that the ice front remained close to the mouth of the Norwegian Channel almost to 18.5 ka.

The smaller readvances mentioned above (Karmøy and Flesen) are probably also represented midway between these

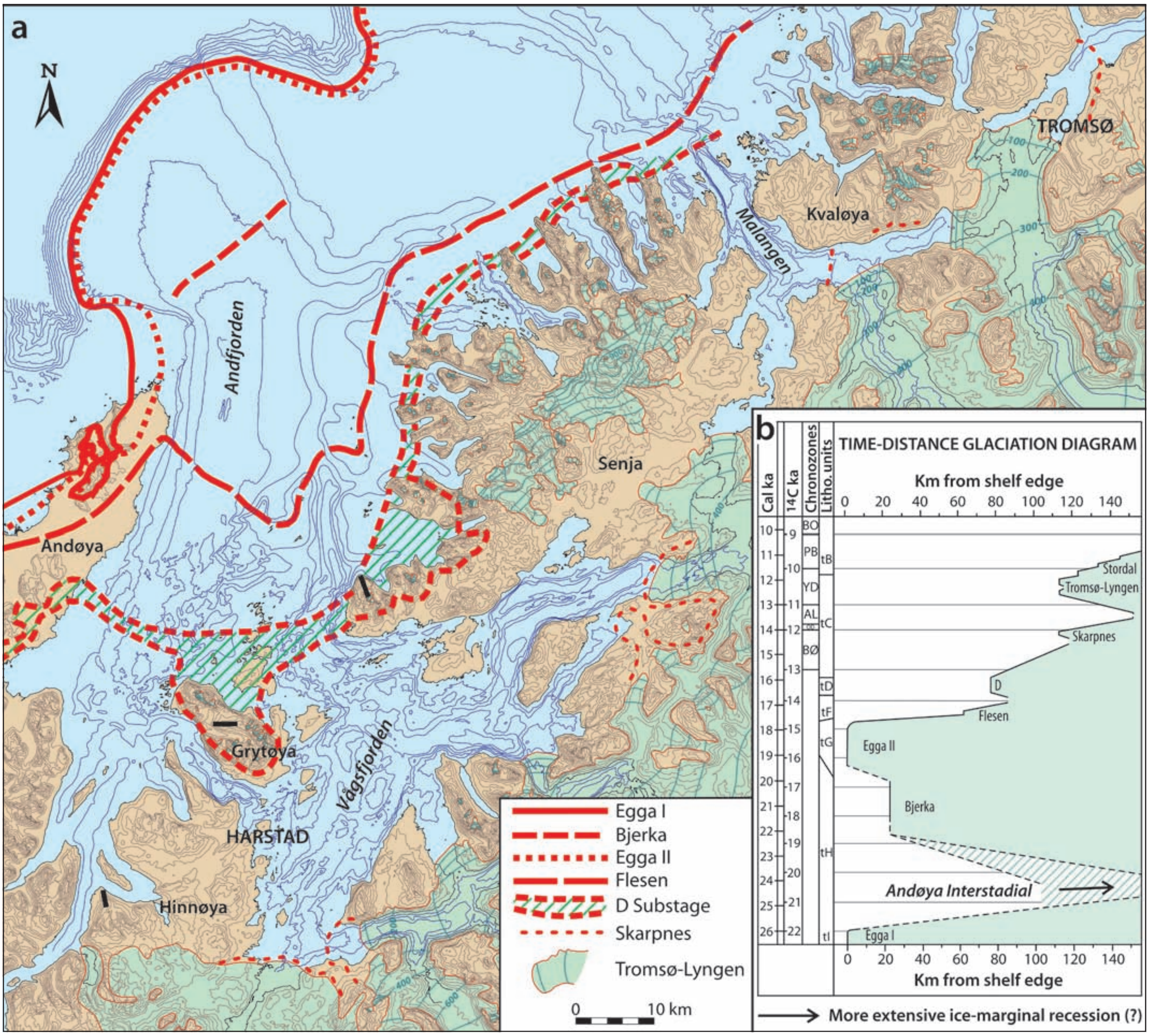


Figure 11. Late Weichselian ice margins (a) and glacial (time–distance) curve (b) showing the ice-front variations near Andøya island, northern Norway. Modified from Vorren and Plassen (2002). The ice margins indicated at Hinnøya in the southwest are after Sveian (2004) and Bergstrøm et al. (2005). Lateral moraines indicated with short black lines at c. 550 m a.s.l., on the southwestern part of Senja, on Grytøya, and possibly also east of Gullesfjord on Hinnøya, are supposed to correlate with the LGM (probably LGM 2; see the main text) that reached to the shelf edge. The ice retreat during Andøya interstadial (c. 24–25 cal ka, i.e., 20–21 ^{14}C ka) is in (b) indicated to reach as far back in the fjords as about 10 km distally to the late-glacial Skarpnes Substage (c. 14.1 cal ka, i.e., 12.2 ^{14}C ka), whereas Olsen et al. (2001a, b, c) have found that this ice retreat probably reached all the way (stippled zone) back to the fjord heads. A revised reconstruction of the glacial variations during LGM, based on new field work on Andøya and new offshore data from the adjoining shelf and fjord areas is now in preparation (T.O. Vorren, pers.comm. 2013). This revision includes, e.g., an adjustment of the age and duration of the Egga II Stadial (the second LGM maximum) to 23-22 cal ka, and renaming of the subsequent interval between 22 and 18 cal ka to Endleten Stadial.

areas, i.e., on the outer part of the Møre–Trøndelag shelf, by the Bremanger and Storegga Moraines (Bugge 1980, King et al. 1987, Nygård 2003). Rokoengen and Frengstad (1999) reported one date of *c.* 18.5 ka from a tongue of till that almost reached the shelf edge in this area, and which probably also correlates with this second readvance.

The conclusion from all available relevant data is that the Late Weichselian ice sheet reached its maximum in western Scandinavia relatively early, probably 28–25 ka. This was followed by a significant ice retreat, in the Andøya–Trofors interstadial, during which the western ice margins receded in most fjord areas (Figure 9) (Vorren et al. 1988, Alm 1993, Olsen 1997, 2002, Olsen et al. 2001a, b, c, 2002). The ice-retreat data include stratigraphical information and many ^{14}C-dated bulk sediments with low organic content. It also includes other dates, e.g., ^{14}C-dated bones of animals and concretions from cave sediments, U/Th-dated concretions from cave sediments, OSL and TL (Optically stimulated luminescence and Thermoluminescence) dated ice-dammed lake sediments (Varangerhalvøya), and most recently also shell dates from Karmøy,

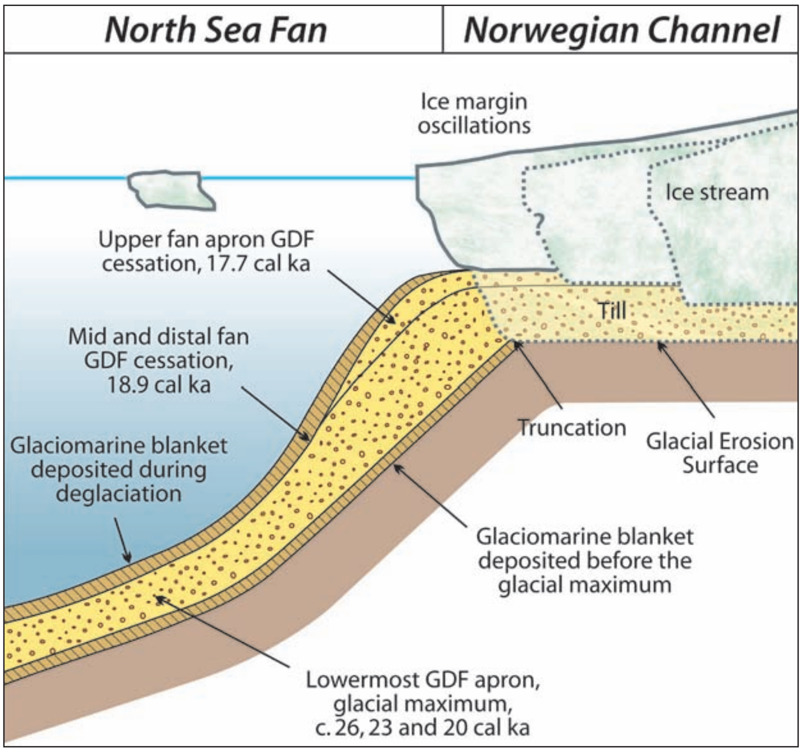


Figure 12. Schematic diagram showing the development of the Norwegian Channel trough-mouth fan during the Late Weichselian glacial maximum. GDF=glacial debris flow. After Mangerud (2004), slightly modified from King et al. (1998), with age assignments for the Last Glacial Maximum from Nygård et al. (2007).

southwest Norway (Figures 3 and 10) (Olsen and Bergstrøm 2007; see also Olsen et al., this issue), as well as OSL-dated subtill glaciofluvial deltaic sediments from Langsmoen, east of Trondheim, Central Norway (Johnsen et al. 2012). Deglaciation sediments, most of these deposited subglacially in the area around Rondane 1000–1100 m a.s.l. (Follestad 2005c), have been OSL dated to *c.* 14–20 ka and found to be well zeroed before deposition (Bøe et al. 2007). The zeroing/resetting of the OSL 'clock' must derive from an earlier depositional phase since no exposure to daylight is supposed to occur during subglacial deposition (this is not discussed by Bøe et al. who assumed subaerial conditions and OSL resetting during final transportation and deposition). Therefore, these dates may represent surficial sediments from the Rondane area during the LGM minimum in the Andøya–Trofors interstadial (Figure 9) and younger nunatak and ice-margin oscillation phases, but which have been subsequently overrun by the ice and later re-sedimented subglacially/sublaterally or at the ice margin during the last deglaciation.

The ice margin in the eastern sector may have continued with advance eastwards, possibly interrupted by minor retreats during the Andøya–Trofors interstadial, from its position during LGM 1 to its maximum position during LGM 2 (Figure 9). Subsequently, at least one major readvance followed (LGM 2) that culminated in the maximum position (shelf edge) or almost the maximum position along most, but not all parts (not on the North Sea Plateau and not in Vesterålen, west of Andøya, Figure 3) of the western flank for the last time at, or shortly after 19.2 ka. However, the ice flows from an ice sheet are time transgressive, with the implication that the maximum position was most likely not reached at the same time everywhere, and for the entire Fennoscandian ice sheet the diachroneity of LGM position seems to be almost 10,000 yr between the western and eastern flanks (e.g., Larsen et al. 1999).

Trough-mouth fans and ice streams on the shelf

During the last two decades, a major contribution to the understanding of glacier dynamics, erosion and deposition on the shelf has come from the record and interpretation of submarine fans at the mouth of glacial troughs ('trough-mouth fans'; Laberg and Vorren 1996, Vorren and Laberg 1997), and the occurrence of distinct ice streams across the shelf during maximum extension of glaciations (see review by Vorren et al. 1998).

The huge North Sea Fan was deposited beyond the Norwegian Channel ice stream (King et al. 1996, 1998, Nygård et al. 2007), which was the longest ice stream on the Norwegian Shelf (Longva and Thorsnes 1997, Sejrup et al. 1996, 1998). This ice stream was 800 km long and drained much of the southern part of the Fennoscandian/Scandinavian ice sheet. In the north, the Bjørnøya Trough ice stream, that also deposited a big fan at the trough mouth, drained much of the southern part of the Barents Sea ice sheet (Vorren and Laberg 1997, Landvik et al. 1998, see also review by Svendsen et al. 2004). In addition, if the Barents Sea and Fennoscandian ice sheets coalesced during maximum extension, as suggested in the 'maximum' model by Denton and Hughes (1981) and in many later reconstructions (e.g., Vorren and Kristoffersen 1986, Landvik et al. 1998, Svendsen et al. 2004, Winsborrow et al. 2010), and both fed the Bjørnøya Trough ice stream, then this ice stream may have drained much of the northernmost part of the Fennoscandian ice sheet.

Detailed morphological mapping of the sea bed has revealed the products of a number of ice streams across the Norwegian shelf (Figures 13 and 14) (Ottesen et al. 2001, 2005a). Numerous large, parallel lineations (megaflutes and megascale lineations) indicating fast ice flow, occur in the troughs, whereas features characteristic of slow-moving ice and stagnant ice are recorded on the shallow areas. However, major trough mouth fans did not form at the outlet of some of these ice streams. Instead of fans, large ice-marginal moraines are located in these positions, and occurring beyond the mouth of the Sklinnadjupet trough west of Sklinnabanken (Figure 13) is the morphologically largest end moraine on the shelf, Skjoldryggen (up to *c.* 200 m high, 10 km wide and 200 km long, Ottesen et al. 2001). For further details from the Norwegian shelf, we refer to e.g., Ottesen et al. (2009).

Ice thickness and ice-surface elevation—did nunataks exist during LGM in Norway?

During the last few decades there has been an increasing understanding of the extreme preservation effect a cold-based ice (frozen to the ground with no sliding) has on its subsurface (e.g., Lagerbäck 1988, Lagerbäck and Robertsson 1988, Kleman

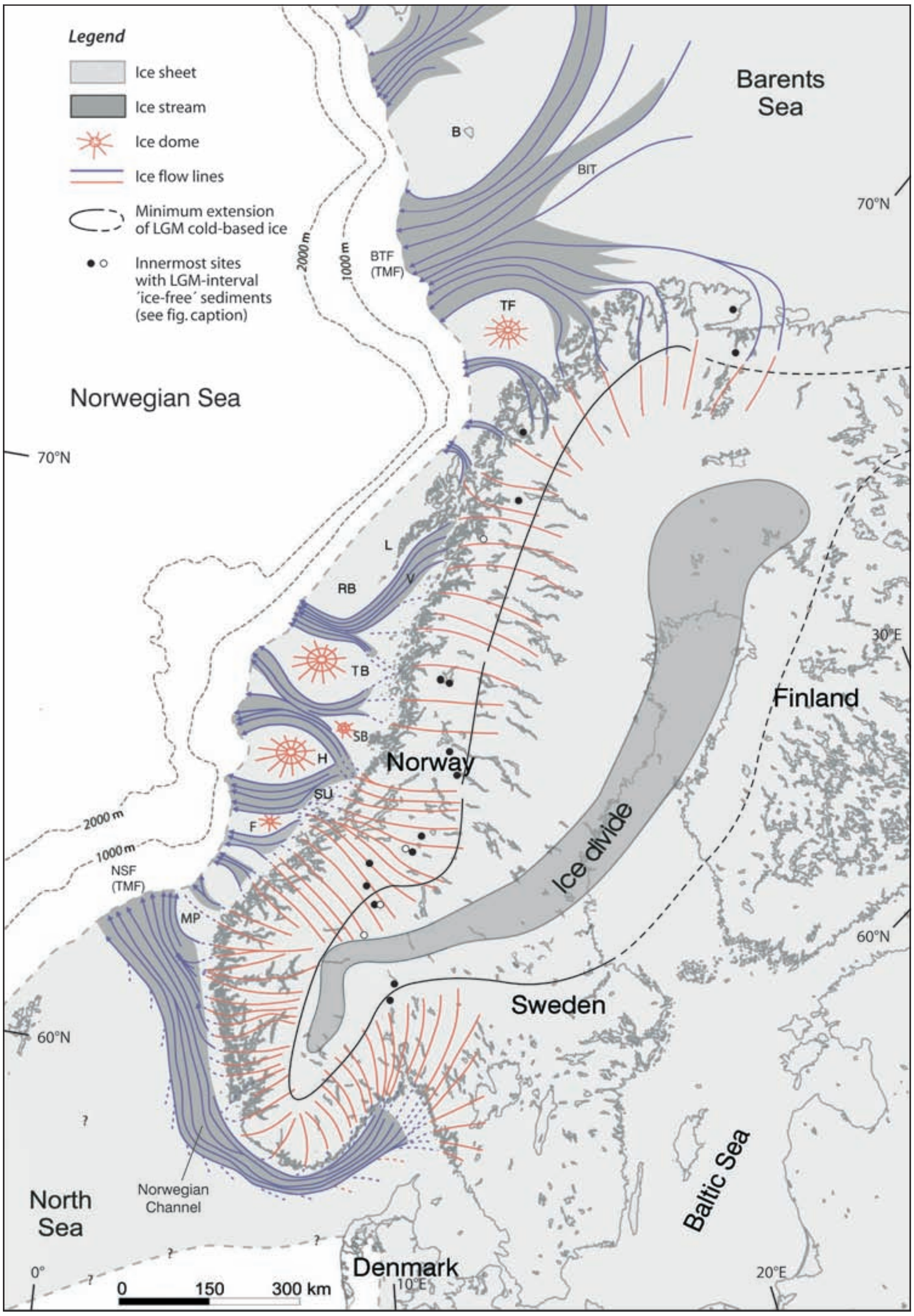


Figure 13. Interpreted ice-flow model during Late Weichselian glacial maximum, from Ottesen et al. (2005a). Minimum areas (discontinuous/stippled line) with ice frozen to the ground (after Kleman et al. 1997) and innermost locations of sites with 'ice-free' sediments from ice marginal retreat during the LGM interval are marked (filled circles indicate stratigraphical data with ^{14}C dates of bulk organics and open circles indicate other dates, mainly after Olsen et al. 2002). B=Bjørnøya (Bear Island), BIT=Bear Island Trough, BTF=Bear Island Trough Fan, TMF=Trough Mouth Fan, TF=Tromsøflaket, L=Lofoten, RB=Røstbanken, V=Vestfjorden, TB=Trænabanken, SB=Sklinnabanken, H=Haltenbanken, SU=Sularevet, F=Frøyabanken, MP=Måløyplatået, NSF=North Sea Fan.

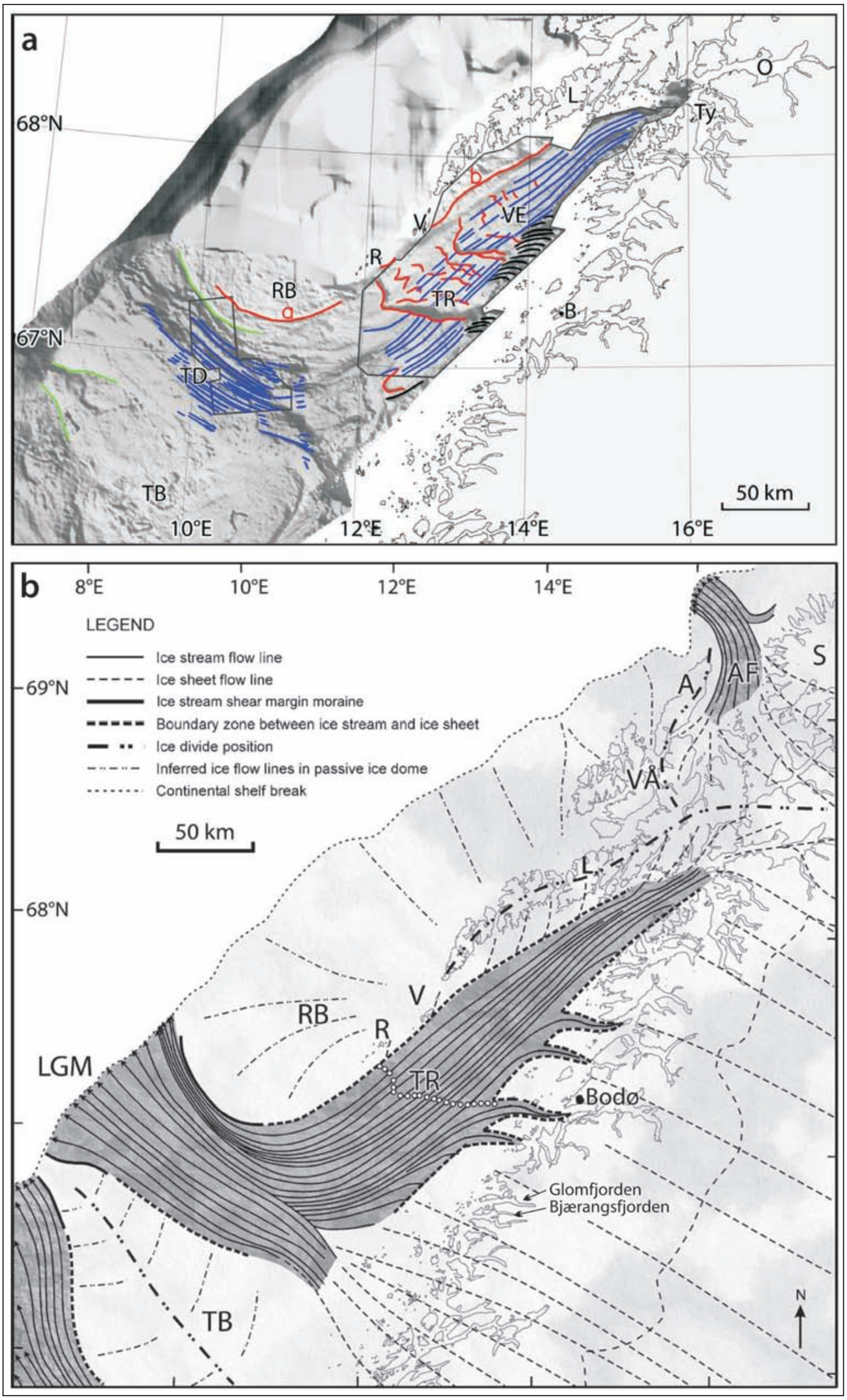


Figure 14. (a) Sea-floor morphology of the Vestfjorden/Trænadjupet with megascale glacial lineation (MSGL) indicating fast ice flow, and (b) Ice flow model of the Lofoten/Vesterålen area and adjacent shelf. Modified from Ottesen et al. (2005b). Legend of letters: O=Ofotfjorden, Ty=Tysfjorden, L=Lofoten, VE=Vestfjorden, V=Værøy, R=Røst, TR=Tennholmen Ridge, B=Bodø, RB=Røstbanken, TD=Trænadjupet, and TB=Trænabanken. Additional legend for (b): S=Senja, AF=Andfjorden, A=Andøya, and VÅ=Vesterålen.

1994, Kleman et al. 1997, Phillips et al. 2006). Therefore, to map which areas that have been ice free versus those which have been covered with a cold-based ice is up to now for most stages an unresolved problem, which imply that the surface geometry, and thus the thickness of the Fennoscandian/Scandinavian ice sheet is poorly known.

A debate, started during the 19th century, on whether mountain peaks in Norway protruded as nunataks above the ice surface during the Quaternary glaciations, especially during the Late Weichselian glacial maximum, is still going on. Mangerud (2004) reviewed this debate and concluded that LGM nunatak areas were possible, and even likely from the Nordfjord area in the south (Figures 15 and 16) to the Lofoten, Vesterålen, Andøya, Senja and Lyngen areas in the north. However, in southeastern Central Norway it seems rather impossible that such nunataks occurred. The LGM ice surface must have reached over most or all mountains in these areas and Mangerud (2004) concluded that most of the authors that he referred to argued that block fields and various other unconsolidated deposits in these inland areas, in most cases, survived beneath cold-based ice. It should also be considered that a pattern of flutes indicating north- to northeastward trending ice flow above 1700–1800 m a.s.l. and up to at least 2100 m a.s.l. in East Jotunheimen east of Sognefjord (Fig.15b; www.norgeibilder.no) shows clearly that the LGM ice sheet must have been thick, and, in places, even partly warm based and erosive up to at least these altitudes. During the last decade, mapping by the Geological Survey of Norway has revealed a record of many glacial accumulation and erosional features in the inland of southeast Norway (e.g., Figure 17). This includes also glaciofluvial lateral drainage channels in some block field areas (e.g., Follestad 2005c, 2006a, b, 2007), which is clear evidence of survival of these block fields underneath a younger cover of a cold-based ice that during final stages produced the meltwater, which channelised the block fields.

It is quite likely that the ice sheet reached its maximum thickness in the west before 26 ka, i.e., during LGM 1 (e.g., Vorren et al. 1988, Follestad 1990a, Møller et al. 1992, Alm 1993, Olsen et al. 2001b, Vorren and Plassen 2002). This is indirectly supported by ice-volume variations on Greenland (e.g., Johnsen et al. 1992, Dansgaard et al. 1993), from onshore areas close to the Vøring Plateau (e.g., Baumann et al. 1995), and also supported by global sea-level data that indicate a maximum sea-level lowering and therefore a maximum global ice volume both late and, even more, early in the LGM interval (Peltier and Fairbanks 2006). It is known that the ice extended to the shelf edge in the west during both LGM 1 and LGM 2, and that the ice extension was roughly the same during these phases both in the north and south. However, the eastward extension was much (c. 500 km) less during LGM 1 than during LGM 2 (Figure 9) (e.g., Larsen et al. 1999, Lunkka et al. 2001), and consequently the area covered by LGM 1 was much smaller than that covered during LGM 2. Therefore, if the volume of the Fennoscandian/

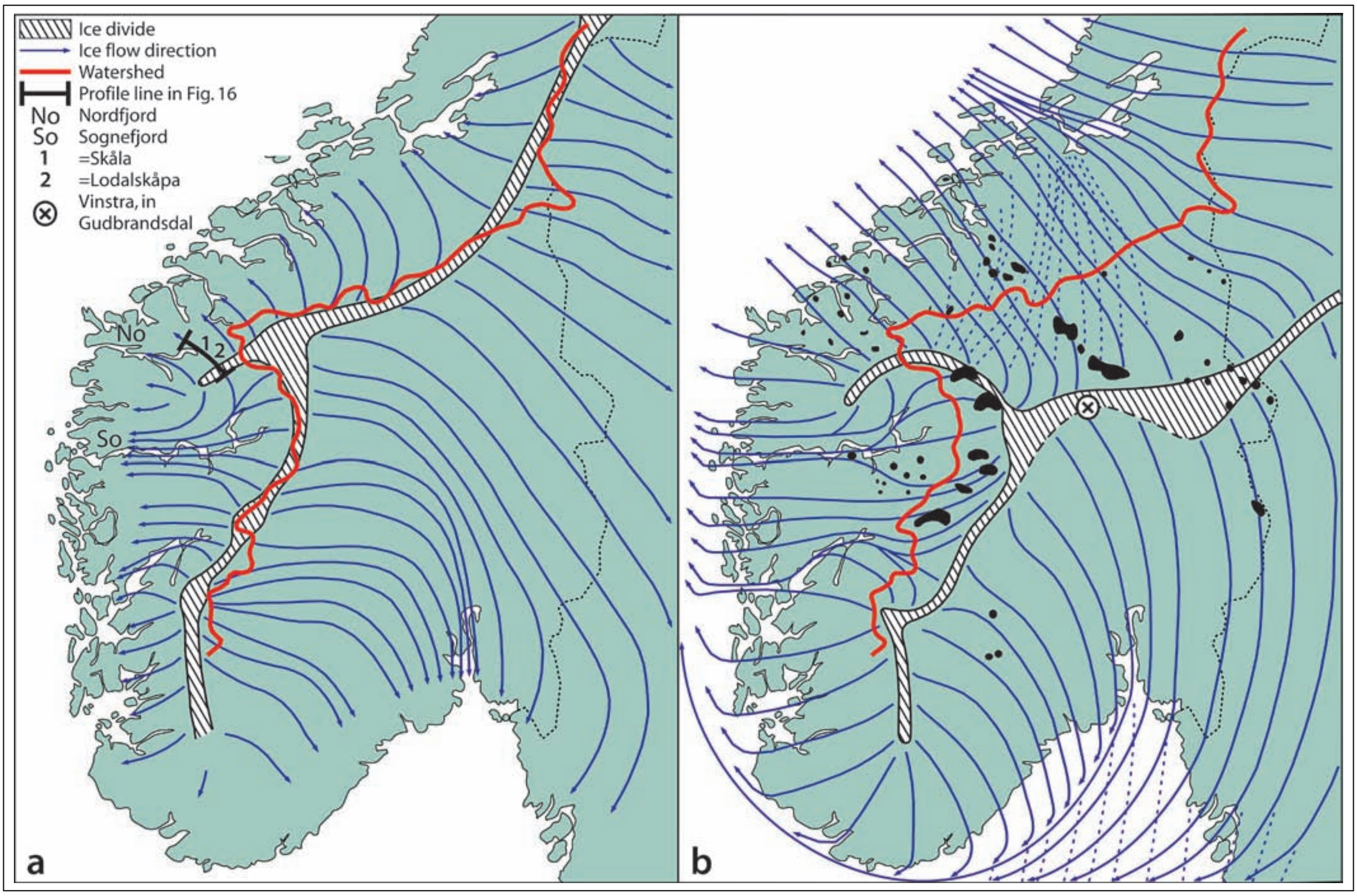


Figure 15. The main flow lines and divides of the ice sheet in southern Norway during small and moderate ice extension (a, left) and the LGM (b, right) are indicated. Modified from Vorren (1979). Blockfields (black spots) are also indicated (map b). Modified from Thoresen (1990). The present authors assume, in accordance with Mangerud (2004), that the cold-based LGM ice sheet covered all the blockfields in the eastern part and possibly also in the west. Stippled flow lines indicate slightly older (and possibly also younger?) LGM ice flows with warm-based ice that reached even to very high altitudes close to the ice shed areas.

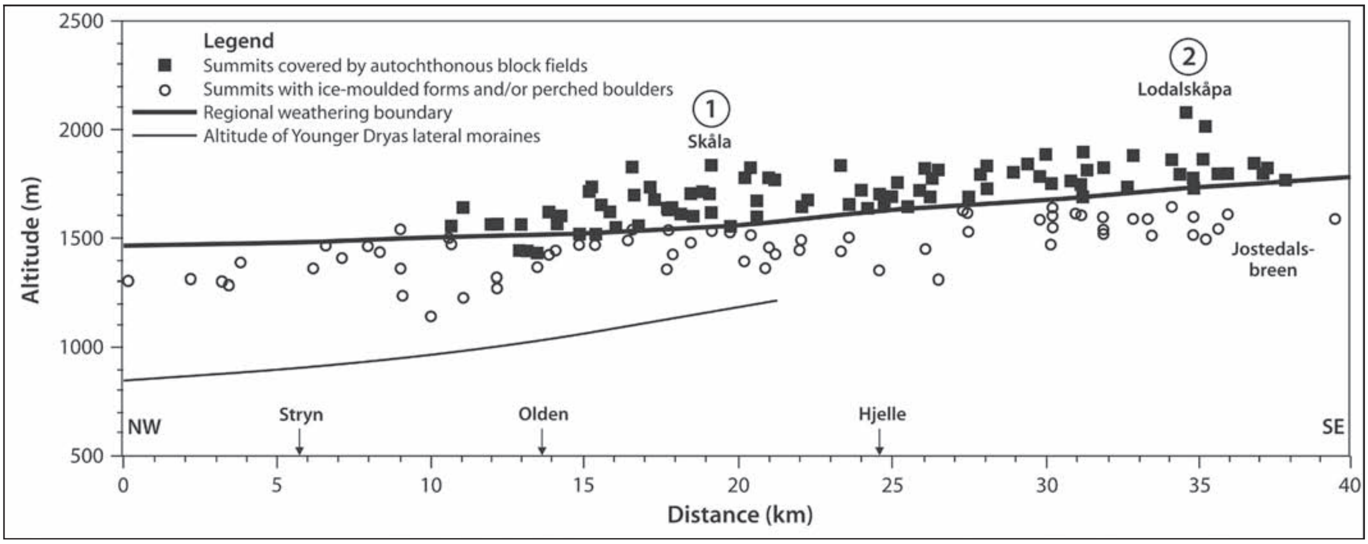


Figure 16. The distribution of summits, with and without block fields are plotted in a NW–SE cross section across inner Nordfjord (see Figure 15). The altitude of the Younger Dryas moraines is also shown. Modified from Mangerud (2004), updated from Brook et al. (1996). The lower boundary of the blockfields is originally assumed to represent the LGM limit, but the present authors assume that the LGM ice sheet may have covered most or all the block fields during LGM 1 (c. 26–27 cal ka, i.e., 22–23 ^{14}C ka) with cold-based ice.

Scandinavian ice sheet follows approximately the overall variations in the ice volume on Greenland and the regional as well as the global ice volumes during the LGM interval, then the ice volumes during LGM 1 was roughly the same as during LGM 2 and, consequently, the ice in the west was probably thickest during LGM 1.

Erratic boulders occurring in terrain with older landforms, such as autochthonous block fields or pre-Late Weichselian deglacial formations are observed by many geologists, including the present authors both in the northern, central and southern parts of Norway. This indicates ice cover and glacial transportation to these fields, but the age of the associated glaciations are in most cases not known. This may soon be changed, because several successful attempts at exposure dating of such boulders have recently been performed in similar settings both in Norway and other places (e.g., Sweden, Canada), and some of these have given late-glacial or early Holocene ages, which indicate a Late Weichselian age for the ice cover and transportation (e.g., Dahl and Linge 2006, Stroeven et al. 2006, Davis et al. 2006).

North Norway

Regional geological mapping indicates total glacial cover for the inland of Nordland and Troms in North Norway, where block fields occur even below the elevation of the Younger Dryas ice surface (Bargel 2003, Sveian et al. 2005). Cold-based ice must also have covered block fields at the Varanger Peninsula in Finnmark, since moraine mounds and glaciofluvial (meltwater) lateral channels are recorded in the block fields, but the age of these mounds and channels, and therefore also the ice cover, is not known (Sollid et al. 1973, Fjellanger et al. 2006). Drumlins, and less common flutings, and striations in very few cases (all representing a sliding warm-based ice), are also recorded within some of the low-lying block fields in Finnmark (Svensson 1967, Malmström and Palmér 1984, Olsen et al. 1996b). This indicates a slightly allochthonous character (glacial input) of parts of these block fields.

Lateral moraines, *c.* 550 m a.s.l. in the Gullesfjord area (on Hinnøya) and at Grytøya and Senja in Troms, that are older than the D Substage and also probably older than the Flesen event (Figure 11) (Vorren and Plassen 2002, Sveian 2004), are supposed to correlate with the LGM, which ended on Andøya and at the shelf edge further north. The ice-surface gradient resulting from this correlation is *c.* 6–9 m km^{-1}, and assuming a similar gradient (*c.* 10 m km^{-1} in the fjord region and slightly less) further inland several possible LGM nunatak areas are recognised (Figure 18), both in high-relief alpine landscapes in fjord areas like Lyngen, but also in some (more) moderate-relief areas along the coast.

The vertical dimensions and timing of the LGM of the Fennoscandian ice sheet in the region northern Andøya to Skånland in northern Norway have been evaluated by Nesje et al. (2007) based on mapping of block fields, weathering boundaries, marginal moraines, and surface exposure dating based on in situ cosmogenic ^{10}Be. They concluded that the LGM ice sheet did not cover the northern tip of Andøya and the adjacent mountain plateaux. Since Nesje et al. (2007) considered the LGM ice thickness only for the period after the significant ice-margin retreat during the Andøya–Trofors interstadial at 22.5–24 ka, their conclusion is valid only for the LGM 2 and not valid for the entire LGM interval, and cold-based ice may well have covered the block fields and the mountain plateaux during LGM 1.

In the area around the present glacier Svartisen in the fjord region of Nordland, North Norway, cosmogenic exposure dating of bedrock (Linge et al. 2007) indicate no erosion during the last glaciation in areas with block fields almost at the same

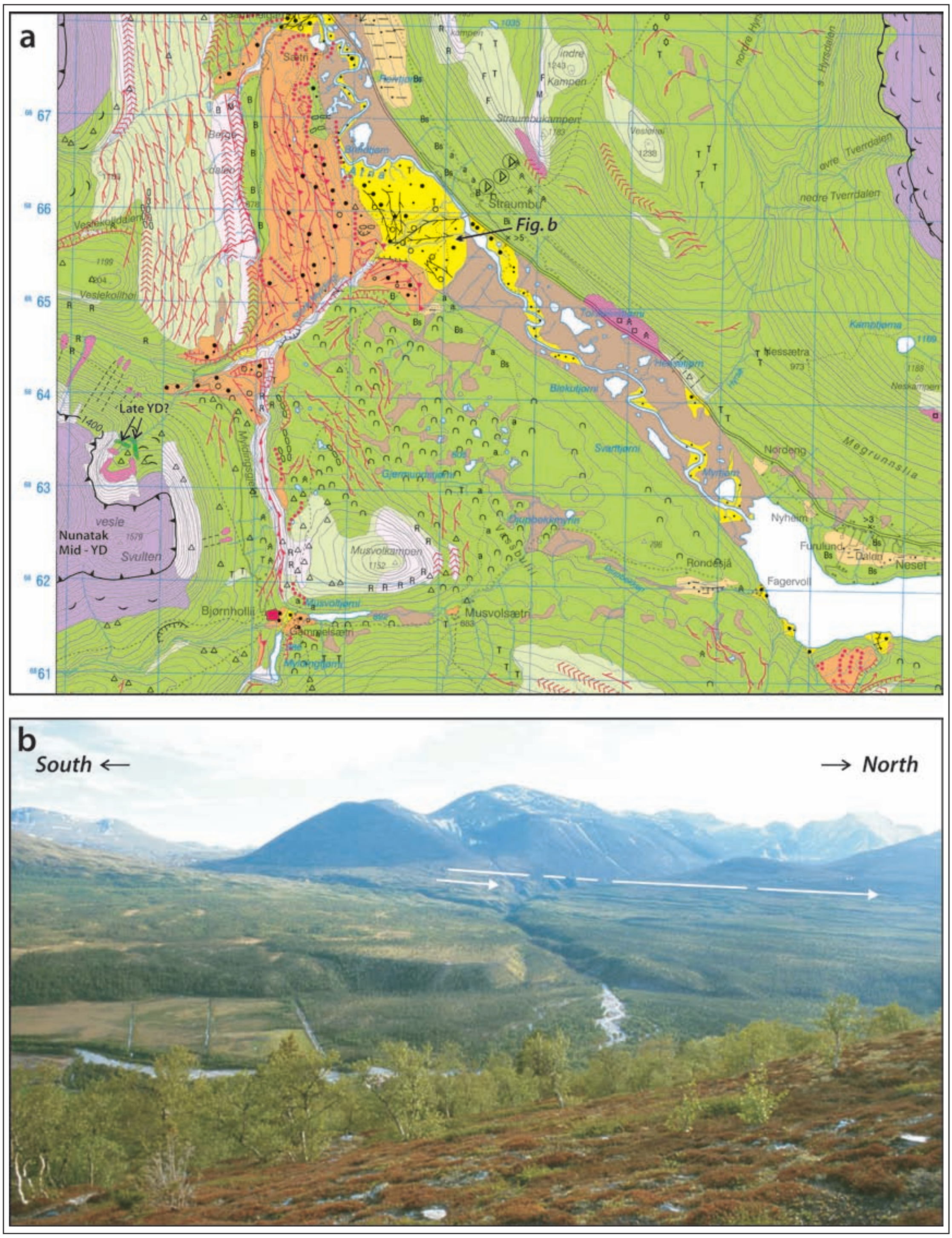


Figure 17. (a) Part of Quaternary map Atnsjøen 1818 IV (1:50,000). The view direction for (b) is marked. Size of each UTM coordinate net square is 1km x 1km. After Follestad (2006a). (b) Lateral meltwater channels from inland southeast Norway. Arrows indicate drainage direction towards the north. Photograph: B.A. Follestad. The delta in the central part is mainly of subglacial origin (see orange field on the map in a). OSL dates from such deposits are not expected to give the age of the last transport and depositional phase, but may rather indicate the age of the source material, in which the OSL signal originally may have been reset to zero.

altitude as the modern glacier surface. The only reasonable explanation for this seems to be that cold-based ice must have covered also these mountains during phases of moderate to maximum glaciation (e.g., during LGM). Similar data is reported from mountain areas of southeastern Norway (Linge et al. 2006), which indicate that cold-based ice on high ground must have been widespread in most parts of Norway during maximum glaciations.

South Norway

At Møre, in the northern part of coastal western Norway many of the high mountains with block fields may have been nunataks during the LGM (e.g., Nesje et al. 1987), but the block fields may also or alternatively have been protected under cold-based ice during a part of the LGM interval as indicated by till fabrics. These data are inferred to represent a thick ice that moved almost topographically independent and possibly reached over the mountains towards the northwest in that area <31 ka

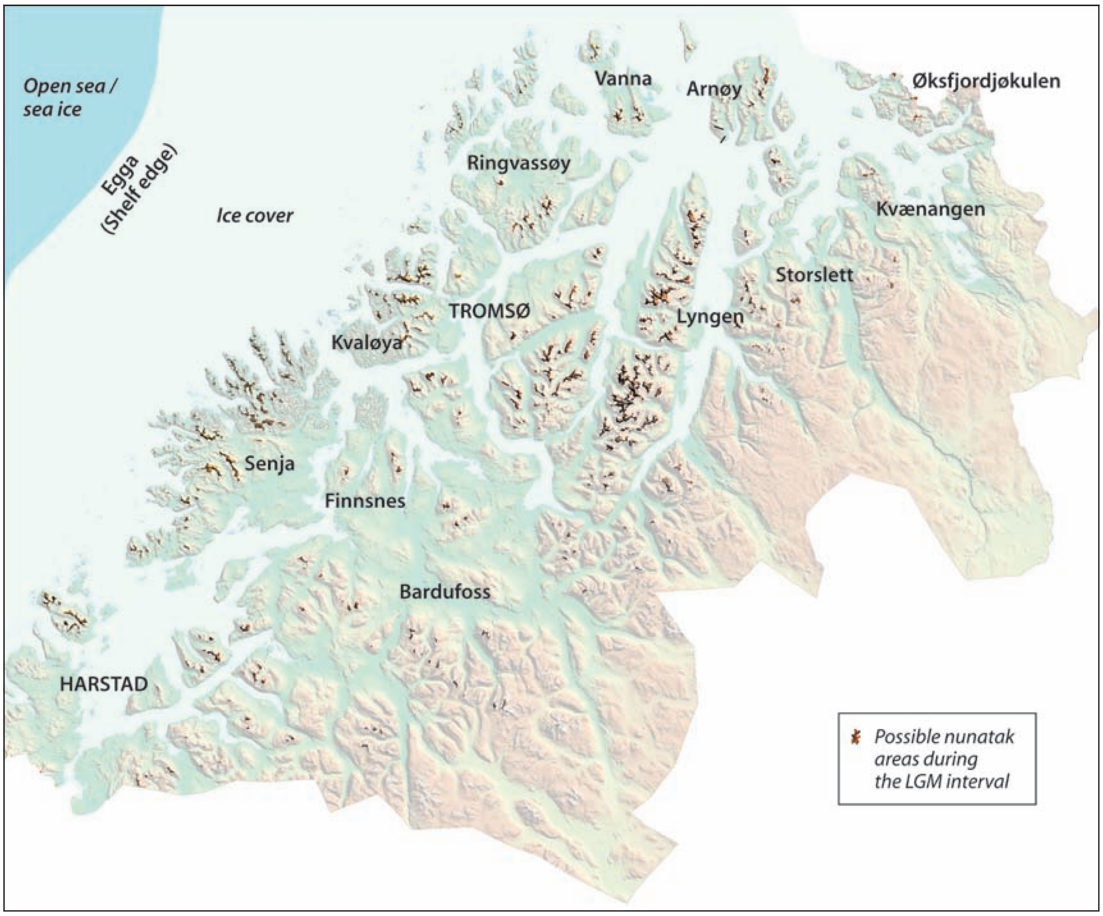


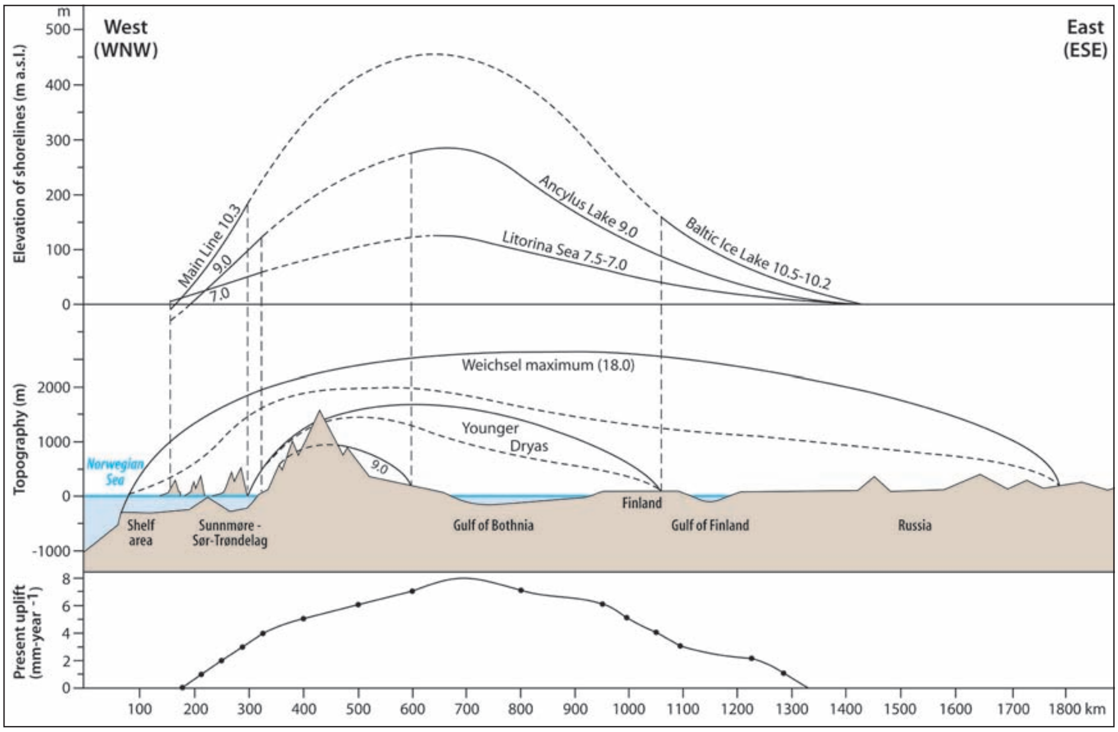
Figure 18. Location of high peaks (brownish shading), e.g., >800 m a.s.l. at Senja, which may have occurred as nunataks and penetrated the LGM ice surface along the coast, similar as in several other high mountain peak areas (>1000–1500 m a.s.l.) in the fjord and inland areas of Troms. Modified from Sveian (2004).

(Follestad 1990a, b, 1992). From the same region, at Skorgenes, strong overconsolidation of clay in subtill position is inferred to result from the load of an ice sheet of late Middle Weichselian or LGM interval age, and that, from estimates of minimum ice thickness, probably covered all the coastal mountains (Larsen and Ward 1992).

A precise level of the maximum LGM ice surface in the ice-divide zone is not known, but roughly estimated values may be considered. For example, the reconstructed Younger Dryas ice-surface positions presented from the region south of Trondheim, reach up to some 1500 m a.s.l. at Røros and Folldal (Sveian et al. 2000, Olsen et al. 2007) and are represented by lateral moraines some 1400–1500 m a.s.l. in Oppdal (Follestad 2005a) and higher than 1500 m a.s.l. in Lordalen south of Lesjaskog (Follestad 2007), which is just below the regional lower level of the block fields (Rudberg 1977, Follestad 2006, 2007). Exposure dating, using the cosmogenic isotope ^{10}Be from a boulder near the summit of Mt. Blåhø at 1617 m a.s.l., several km north of the ice divide during the LGM, yielded an age of 25.1 ± 1.8 ka, which suggests that Mt. Blåhø was entirely covered by ice during LGM (Goehring et al. 2008). This indicates that the average LGM ice surface must have been at least 2000 m a.s.l. at the ice divide. This is a source area for the ice flowing via Oslofjorden to Skagerrak and further feeding the 800 km-long Norwegian Channel ice stream. With a gradient as low as 1 m km^{-1} for the ice stream and 7 m km^{-1} as average gradient for the remaining part (200 km) northwards to the ice divide, the LGM ice surface at the ice divide would reach at least 2100 m a.s.l. Compared with modern analogues from Greenland and Antarctica it seems rather clear to the present authors that this is a minimum estimate. In addition, the western part of the ice-divide zone, the Jotunheimen area, where also the highest mountains (maximum 2469 m a.s.l.) are located, seems to have been a dome area also during the LGM interval and the ice surface would therefore have been even higher there.

Longva and Thorsnes (1997) described three generations of ice-flow directions based on the sea-bed morphology at a plateau south of Arendal, northern Skagerrak. The oldest generation was represented by deep diffuse furrows with a direction showing that the ice crossed the Norwegian Channel. This ice flow was considered to reach the Late Weichselian ice maximum in northern Denmark (Longva and Thorsnes 1997), which is supported by all relevant recent studies in Denmark (e.g., Houmark-Nielsen 1999, Houmark-Nielsen and Kjær 2003). Deep furrows, with a more southwesterly direction, represent the next generation. This change in direction was probably a result of the well-known eastward migration of the ice-divide zone over Central Scandinavia during the LGM interval. The youngest ice flow is represented by flutes from a plastic ice movement along the Norwegian Channel, and is interpreted to

Figure 19. Profiles across Central Fennoscandia, generalised and modified from various sources, including Svendsen and Mangerud (1987), Mangerud (2004) and Påsse and Andersson (2005). Ages in ¹⁴C ka. Profile line indicated in Figure 1. Upper panel: *shore lines of different ages.* Middle panel: *alternative ice-sheet profiles for the Late Weichselian maximum (18–15 ¹⁴C ka) and the Younger Dryas maximum. Full lines show maximum thickness, stippled lines minimum thickness.* Lower panel: *present-day uplift.*



represent an ice stream in the channel (Longva and Thorsnes 1997).

The referred observations from northern Skagerrak support a hypothesis that the LGM developed as a thick Scandinavian ice sheet, initially from a westerly located core area, e.g., as the one presented by Kleman and Hättestrand (1999). During this ice phase there was no ice stream in the Norwegian Channel. Ice from Norway could then move and transport Norwegian erratics across the Norwegian Channel to Denmark and the North Sea. In Denmark and the shallow part of the North Sea, there was probably permafrost during the advance, favouring a steeper ice surface (Clark et al. 1999). All summits in South Norway could in this early phase have been covered by cold-based ice. Subsequently the ice streams developed, probably from the shelf edge and migrating upstream, which would lead to a considerable downdraw of the ice-sheet surface, and ending with a situation similar to the western part of that reconstructed by Nesje et al. (1988) and Nesje and Dahl (1990, 1992), but with a significant deviation from their model in the eastern part. In this area they did not account for a change of position for the LGM ice divide compared to phases with smaller ice extension. In this central inland region we suggest that the ice surface have reached to higher relative elevations, probably covering most or all block fields also after downdraw, and the ice-divide zone migrated to a much more southerly and southeasterly position, as indicated in Figure 15. The eastwardly migrating ice divide was hypotesised by Vorren (1977), who explained this as a result of downdraw after major surges in the fjords at the western ice margins. Reconstructions based on field data, including till stratigraphy (e.g., Bergersen and Garnes 1981, Thoresen and Bergersen 1983, Olsen 1983, 1985, 1989) have subsequently supported the idea of an eastwardly migrating ice divide.

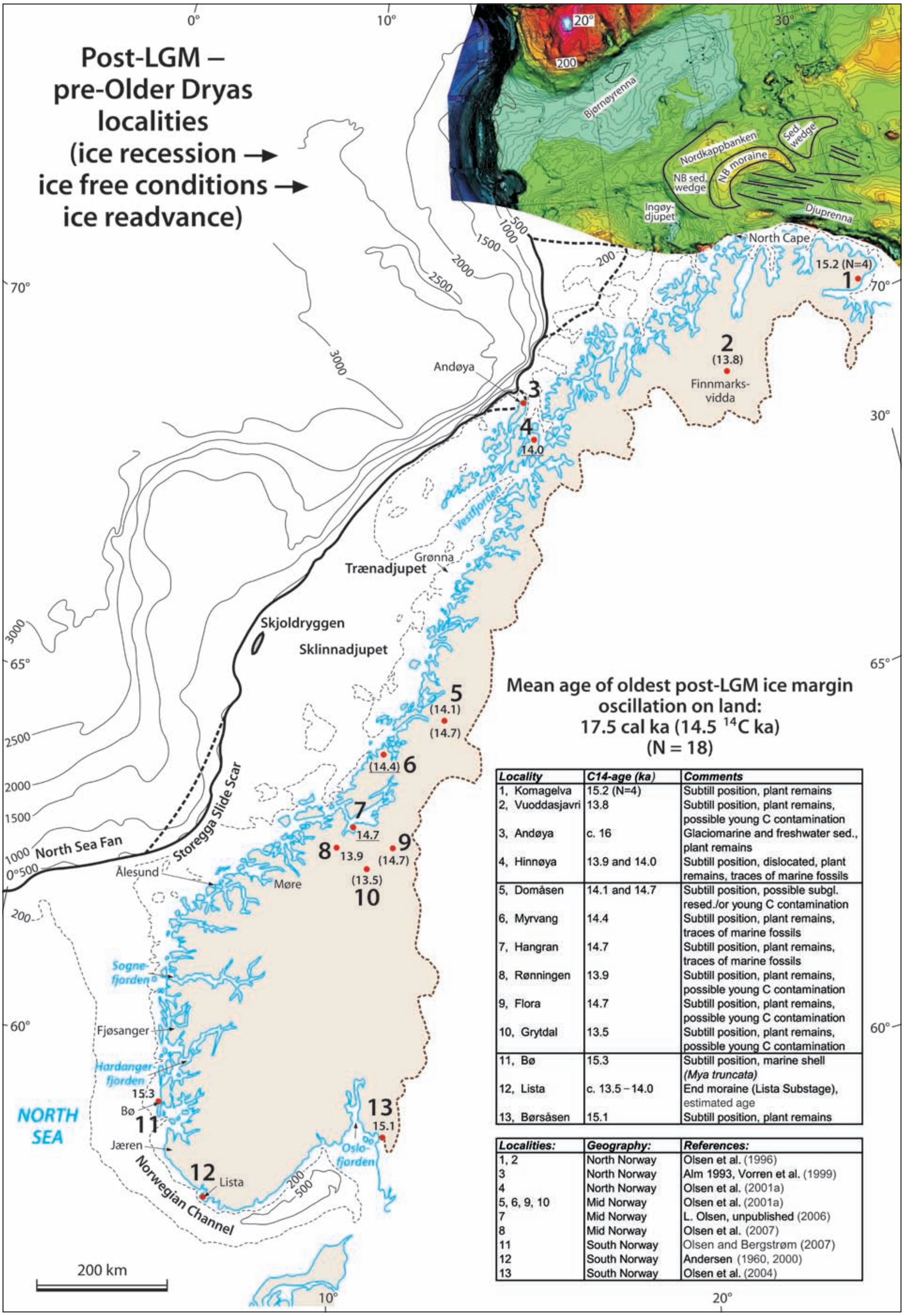
The same result with downdraw and changed geometry of the ice sheet would also be expected if the ice streams developed with initiation in the proximal parts (e.g., in the Oslofjord–Skagerrak area), where big subglacial water bodies, if they existed may have been suddenly drained and caused a considerable increase in the ice-flow velocity. Modern analogues from Antarctica (e.g., Bell et al. 2007) suggest that ice-stream triggering like this should be considered.

The surface geometry and thickness of the Scandinavian ice sheet are indicated by E–W cross profiles across the central area (Figure 19). However, generally lower ice-surface gradients are assumed to have existed in areas where the ice has slid on the substrate or moved forward on deformable beds (e.g., Olsen 2010).

The oldest post-LGM ice-marginal moraines on land

The locations of the oldest post-LGM deglaciation sediments on land with indication of subsequent ice advance over the site are shown in Figure 20. The Late Weichselian Karmøy/Bremanger ice readvance occurred shortly after 18.6 ka and reached off the coast in southwest Norway (Nygård 2003, Nygård et al. 2004, Olsen and Bergstrøm 2007). The ice margin may have retreated to onland positions and readvanced around 18.5 ka in the outer coastal areas of northern Norway. The ice-margin continuation on Andøya of the ice-front formations representing the Flesen event, which is recorded in the adjacent Andfjorden area (Vorren and Plassen 2002) is here considered to possibly correlate with the Late Weichselian Karmøy/Bremanger readvance.

The Risvik Moraines in Finnmark (Figure 21, Sollid et



Locality	C14-age (ka)	Comments
1, Komagelva	15.2 (N=4)	Subtill position, plant remains
2, Vuoddasjavri	13.8	Subtill position, plant remains, possible young C contamination
3, Andøya	c. 16	Glaciomarine and freshwater sed., plant remains
4, Hinnøya	13.9 and 14.0	Subtill position, dislocated, plant remains, traces of marine fossils
5, Domåsen	14.1 and 14.7	Subtill position, possible subgl. resed./or young C contamination
6, Myrvang	14.4	Subtill position, plant remains, traces of marine fossils
7, Hangran	14.7	Subtill position, plant remains, traces of marine fossils
8, Rønningen	13.9	Subtill position, plant remains, possible young C contamination
9, Flora	14.7	Subtill position, plant remains, possible young C contamination
10, Grytdal	13.5	Subtill position, plant remains, possible young C contamination
11, Bø	15.3	Subtill position, marine shell *(Mya truncata)*
12, Lista	c. 13.5 – 14.0	End moraine (Lista Substage), estimated age
13, Børsåsen	15.1	Subtill position, plant remains

Localities:	Geography:	References:
1, 2	North Norway	Olsen et al. (1996)
3	North Norway	Alm 1993, Vorren et al. (1999)
4	North Norway	Olsen et al. (2001a)
5, 6, 9, 10	Mid Norway	Olsen et al. (2001a)
7	Mid Norway	L. Olsen, unpublished (2006)
8	Mid Norway	Olsen et al. (2007)
11	South Norway	Olsen and Bergstrøm (2007)
12	South Norway	Andersen (1960, 2000)
13	South Norway	Olsen et al. (2004)

Figure 20. Oldest post-LGM ice-oscillation sites on land in Norway. Inset map indicates glacial features on the sea bed of the southwest Barents Sea, modified from Andreassen et al. (2008). Note the position of the post-LGM Nordkappbanken arcuate moraine.

al. 1973, Olsen et al. 1996b) may be of the same or a slightly younger age, but the glacier dynamics seem to be different since these moraines are not considered to represent one distinct regional readvance. They merely represent a series of halts and local readvances during overall ice recession after the major offshore Nordkappbanken Substage (Figure 20), which is represented by a large arcuate moraine that marks the front of an ice lobe (Andreassen et al. 2008), which may correlate with the Late Weichselian Karmøy/Bremanger readvance. These correlations imply that the mean value of five dates from subtill lake sediments, that gave a maximum age of 19.7 ka for the Risvik Substage on Varangerhalvøya (Table 1), is too high by at least some 1300–1500 years. We suggest, in accordance with data from previous studies by e.g., Hyvärinen (1975, 1976), Prentice (1981, 1982) and Malmström and Palmér (1984), that an inherited ('reservoir') age of this size (1000–2000 years) may possibly be represented in carbonate-rich waters in lacustrine basins from shortly after the last deglaciation on Varangerhalvøya.

A similar consideration may be given for the subsequent ice-marginal zone on Varangerhalvøya, the Outer Porsanger/Vardø Substage (Figure 21), with a mean value of three dates of subtill lake sediments yielding a maximum age of 18.7 ka (Table 1). A 'reservoir' age of approximately the same size as suggested above gives apparently a more correct age of *c.* 16.7 ka for this substage. The chronology based on shoreline data from Finnmark (Marthinussen 1960, 1974, Sollid et al. 1973) has a resolution that is too low to give precise ages of the oldest ice-marginal substages. Based on the considerations above, the mentioned shoreline data, and on regional correlations of younger ice-marginal substages in Finnmark (Figure 21), we support the proposal by Andersen (1979) of using preliminary ages of *c.* 18.5 ka for the Risvik Substage and *c.* 16.7 ka for the Outer Porsanger/Vardø Substage. However, recently reported dates at 15 ka for marine shells and algae from basins at Magerøya and Nordkinn, in the zone of the Risvik Substage after Sollid et al. (1973), indicate that the suggested ages of the Risvik and Outer Porsanger Substages may be 1000–2000 years too old (Romundset 2010).

A large morainal bank, possibly a grounding-line moraine (30–35 m high, 1–2 km wide and *c.* 10 km long) which crosses Porsangerfjorden about 10–20 km from its mouth (Ottesen et al. 2008), seems to correspond with the suggested seaward extension of the Risvik Substage in this area (Olsen et al. 1996b). Another large moraine that crosses Porsangerfjorden some 15–20 km farther south corresponds with lateral moraines (c. 300 m a.s.l.) of the Outer Porsanger Substage some 10–20

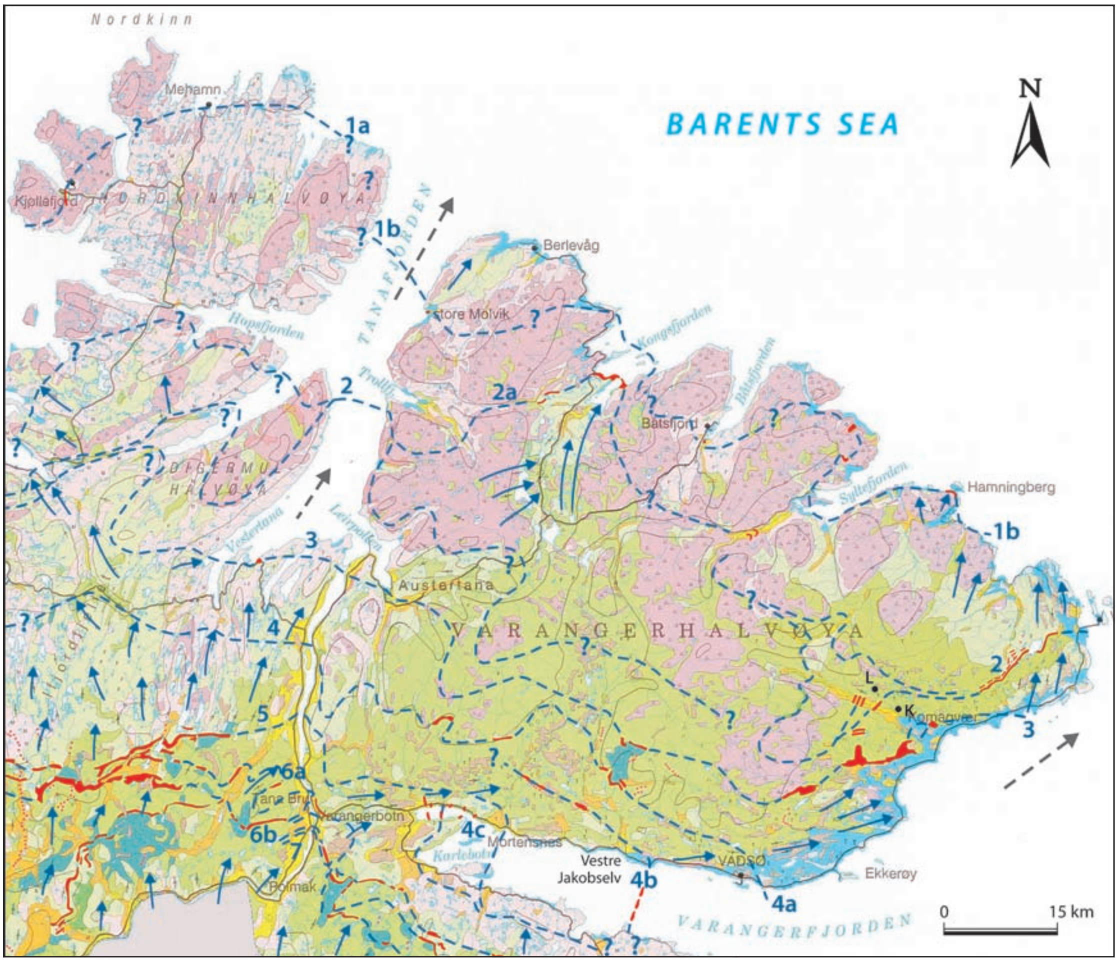


Figure 21. Inferred ice margins (1a, 1b, 2a, 2, 3, 4a, 4b, 4c, 5, 6a, 6b and 7) in northeast Finnmark, northern Norway during the Late Weichselian. Modified from Sollid et al. (1973) and Olsen et al. (1996b). 1a, 1b= The Risvik Substage, and 2, 2a= The Outer Porsanger - Vardø Substage.

Table 1. Dates (^{14}C and OSL) which may represent the Late Weichselian Risvik and Outer Porsanger (O.P.) ice-marginal substages, from Leirelva and Komagelva sites on Varangerhalvøya, northeast Norway. Dates are from Olsen et al. (1996a).

Risvik Substage; Leirelva		O. P. Substage; Komagelva		
No.	*^{14}C-age*	*No.*	*OSL-age*	^{14}C-age
1	15.0 ka	1		15.4 ka
2	17.1 "	2		16.4 "
3	17.3 "	3	17.0 ka	c. 14.5 ka
4	18.7 "	4		14.4 ka
5	14.6 "			
	16.5 ^{14}C ka			15.4 ^{14}C ka
Average age;	19.7 cal ka			18.7 cal ka

A 'reservoir' age from late-glacial and early post-glacial lacustrine environments at Varangerhalvøya is assumed to be 1300–1500 ^{14}C years (see the main text). Adjusted maximum ages for the Risvik and O.P. Substages are therefore c. 18.5–18.6 cal ka (15.0–15.2 ^{14}C ka) and 16.5–16.8 cal ka (13.9–14.1 ^{14}C ka), respectively.

km farther south along the fjord (Olsen et al. 1996b). This gives an average ice-surface gradient of *c.* 15 m km^{-1} along the last 10–20 km towards the ice front. The dimension of the fjord-crossing moraine and its supposed continuations on both sides of Porsangerfjorden and farther east in Finnmark indicate a marked regional readvance of the ice sheet during this substage, which is different from the glacial dynamics during the Risvik Substage. Moraines in the outermost coastal parts of Troms may be of the same age as the Risvik and/or the Outer Porsanger Substages, but none of these are dated yet (Andersen 1968, Sveian et al. 2005).

The ice margin retreated onshore at Jæren in southwest Norway probably between 17 and 16 ka or before 16.7 ka (Thomsen 1982, Paus 1989, Matthiasdóttir 2004, Knudsen et al. 2006), and the ice-marginal moraines some 6 km inland from the present coastline southwest of lake Storamos may have an age of *c.* 16.7 ka (Andersen et al. 1987, Wangen et al. 1987, Matthiasdóttir 2004). At Lista, farther south (Figure 20), the oldest post-LGM ice-marginal moraine (the Lista Substage) follows the outermost coastline for several kilometres, and has an estimated age of *c.* 16.1–16.7 ka (Andersen 1960, 2000). Field observations by the Geological Survey of sections in the Lista moraine in the year 2010 suggest that structural influence (deformation, redeposition, possible new input of rock material from the Oslo region) from a possible coexisting ice body in the Norwegian Channel may have occurred. This indicates a complex genesis of the Lista moraine, which may be a combined feature formed at the ice margins between the terrestrial ice flowing towards and ending at the margin laterally to the Norwegian Channel and an ice stream moving and making a lateral shear moraine along the channel.

In the outermost coastal areas between Jæren in the south and Andøya–Senja in the north there are scattered moraines that may be older than 15.3 ka, but most of these are not dated yet. One exception is the large grounding-line moraine, the Røst moraine (Tennholmen Ridge), that has been mapped in Vestfjorden (Rokoengen et al. 1977, Ottesen et al. 2005b). It has its continuation on Grønna and other small islands/skerries about 20–25 km northwest of Meløya in mid Nordland (Bøe et al. 2005, 2008) and is dated as part of the Røst marginal-moraine system with an age limited between 16.3 and 15.5 ka (Knies et al. 2007, Laberg et al. 2007).

Local cirque moraines at *c.* 200 m a.s.l. that are recorded in the southwestern part of Andøya have recently been suggested to derive from glacier activity between 21 and 14.7 ka (Paasche et al. 2007). This suggestion is hampered by a lack of dating support, and even with an age model like the one they have used, and which seems quite reasonable, a late-glacial age cannot be disproved. Therefore, we consider the age of these moraines at present to be undetermined.

The Lateglacial period

To clarify the terminology, in the following the Lateglacial period represents the time between 13 and 10 ^{14}C ka (e.g., Ehlers 1996). The Lateglacial is in northern Europe subdivided into the Lateglacial Interstadial and the Younger Dryas Stadial (Berglund et al. 1994). However, a subdivision of the Lateglacial Interstadial in the classical intervals, the Bølling interstadial (15.3–14.1 ka), the Older Dryas (OD) stadial (14.1–13.8 ka, suggested age interval after Olsen 2002) and the Allerød interstadial (13.8–12.9 ka), are maintained for the Norwegian areas since these intervals/events, particularly the Older Dryas stadial (glacier readvance), are well expressed in the geological record of Norway, as reviewed by e.g., Olsen (2002). It should also be added that the term 'interstadial' is here used for phases of local ice-free conditions, significant ice retreats and glacier minima rather than merely based on temperature and vegetation criteria, which is the standard use in areas farther from the ice margins.

Dating problems

There are five main problems with ^{14}C dates that are relevant for the accurate dating of Lateglacial glacial events in Norway. First, the calibration to calendar years is still not well established for the pre-Holocene, and particularly pre-LGM ages (however, calibration for older ages is significantly improved during the last decade). Second, there are plateaux in the radiocarbon scale so that intervals in calendar years, rather than specific, precise ages represent certain ^{14}C ages (i.e., a conversion from ^{14}C to calendar years, or vice versa do not follow a simple linear function). Third, among the previously dated marine molluscs there are different types of species; some feed by filtering of particles from seawater, others are surface-sediment feeders and a third group of species belongs to the subsurface feeders. The risk of contamination by old carbon is obviously much higher for sediment feeders than for filter feeders (Mangerud et al. 2006). Fourth is the problem of possible reworking of older microfossils, such as foraminifera, by currents. Fifth is the uncertainty related to the marine reservoir age, because most ^{14}C dates of the Lateglacial moraines in Norway are performed on marine molluscs.

Uncertainties in the reservoir age hamper precise comparison between dates on marine and terrestrial materials (Mangerud 2004). Conventionally all dates of marine fossils from the Norwegian coast are corrected for a reservoir age of 440 years (Mangerud and Gulliksen 1975). However, it is now known that the marine reservoir age for western Norway was *c.* 380 years for the Allerød (Bondevik et al. 1999), increased to 400 years during late Allerød and further to 600 years in the early Younger Dryas, stabilised for 900 years, and dropped to 300 years across the Younger Dryas–Holocene transition, and is today 360 ± 20 years (Bondevik et al. 2006). If these values are correct for all parts of Norway, then this has significant regional implications. For example, some moraines that have during the last decades been considered to be of early to middle Younger Dryas age could be of late Younger Dryas age, and moraines assumed to be of early late Younger Dryas age or very early Preboreal age, based on mollusc dates, might in fact be of early Preboreal age or late Younger Dryas age, respectively.

Figure 22. The Younger Dryas moraines around Fennoscandia, slightly modified from Andersen et al. (1995) and Mangerud (2004). The names of the moraines are given in bold letters. Other geographical names used in the discussion of the Younger Dryas glacial limit are shown with normal text.

Diachronous moraines and significant regrowth of ice

End moraines from the Younger Dryas have been mapped more or less continuously around the entire former Scandinavian ice sheet (Figure 22) (as reviewed by Andersen et al. 1995), and in most parts of Norway the Older Dryas end moraines (c. 13.8 ka) are also well represented (Andersen 1968, Sollid et al. 1973, Andersen et al. 1979, 1982, 1995, Sørensen 1983, Rasmussen 1984, Follestad 1989, Olsen et al. 1996b, Sveian and Solli 1997, Olsen and Sørensen 1998, Bergstrøm 1999, Olsen 2002, Sveian et al. 2005, Olsen and Riiber 2006). However, it is clear that the outermost and largest Younger Dryas moraines are diachronous around Fennoscandia (Mangerud 1980), although most of these moraines were formed during the early and middle part of Younger Dryas. The zone that deviates most from the general trend is southwest Norway, and particularly the Hardangerfjord–Bergen area, where the ice sheet readvanced during the entire Younger Dryas, and formed the Halsnøy and Herdla moraines at the very end of the Younger Dryas (Appendix B, Figure B4) (Lohne et al. 2007a, b). The Older Dryas moraines are also apparently diachronous around the coast of Norway, although most dated moraines from the Older Dryas readvance are *c.* 13.8 ka and always between 14.2 and 13.7 ka (Andersen 1968, Rasmussen 1981, 1984, Sørensen 1983, Follestad 1989, Larsen et al. 1988, Sveian and Olsen 1990, Bergstrøm et al. 1994, Bergstrøm 1999, Olsen 2002, Olsen and Riiber 2006).

Bølling interstadial–Older Dryas readvance

The climatic amelioration that initiated the Bølling interstadial *c.* 15.3 ka in Norway lead to significant but still not well recorded ice retreat in the fjord regions. Some dates of plant remains from subtill sediments indicate that the ice retreat may have reached back to the fjord valleys of Mid-Norway during the Bølling interstadial (Kolstrup and Olsen 2012), but dates of marine shells of this age are so far only represented from the coastal zones in most parts of Norway. However, exceptions exist and one of these is found in the area around the Arctic Circle, where shell dates of Bølling age occur at several sites in the fjord region (Olsen 2002). The locations of the ice margin before the Older Dryas and Younger Dryas readvances are not precisely known, but generally the Older Dryas ice advanced a few kilometres downstream. For example, in Trøndelag, Mid-Norway, an Older Dryas readvance of at least 5–10 km is recorded (Olsen and Sveian 1994, Sveian and Solli 1997, Olsen and Riiber 2006), and in northern Norway, west of the Svartisen glacier, an ice advance of at least 10–15 km is estimated for the Older Dryas (Appendix B, Figure B5b) (Olsen 2002), which is the maximum reported Older Dryas readvance in Norway.

Allerød interstadial–Younger Dryas readvances

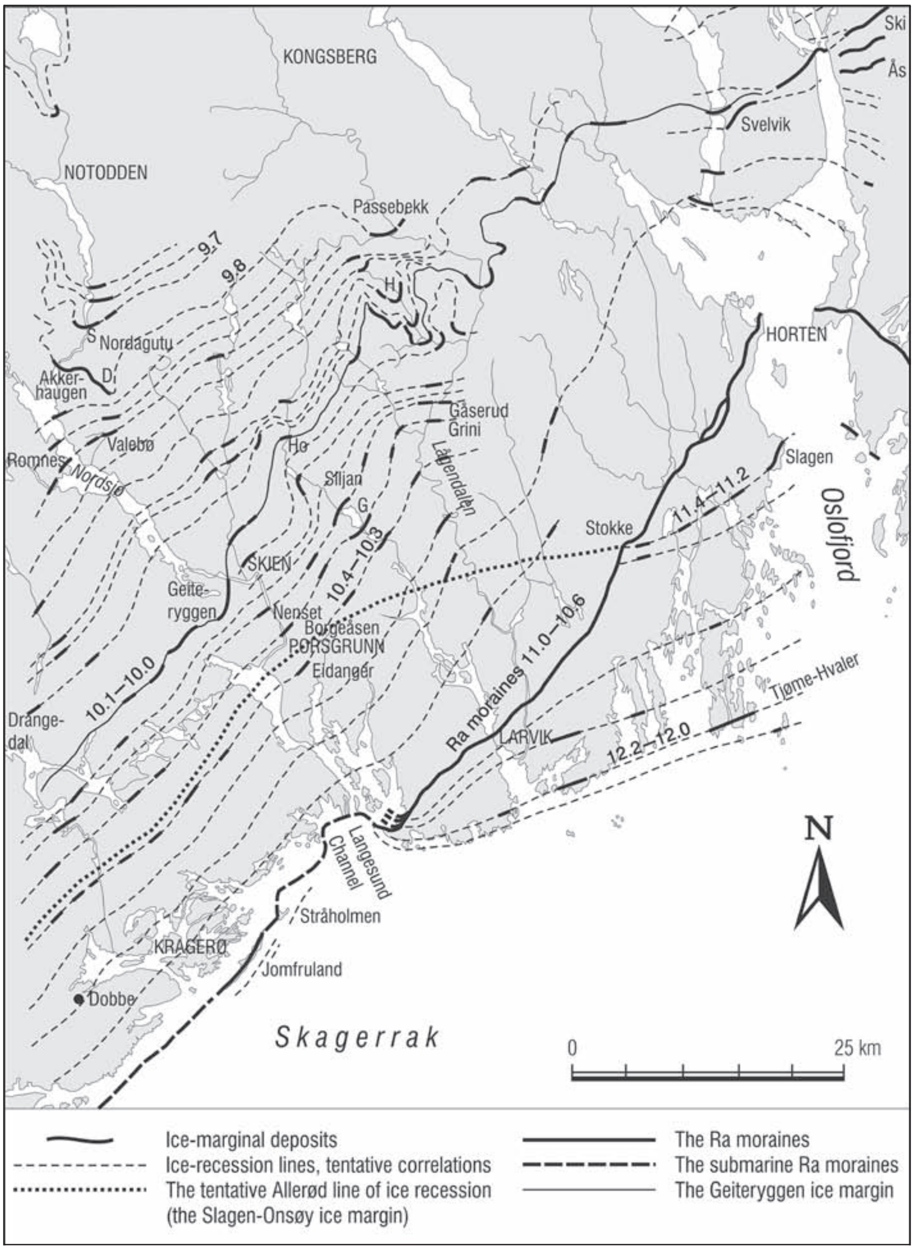
The ice retreat during the Allerød interstadial is generally assumed to have reached the heads or innermost parts of most fjords in Norway. Dates of marine shells of Allerød age and stratigraphical evidence supports this interpretation for southern Troms (Vorren and Plassen 2002, Bergstrøm and Olsen 2004), Ofotfjorden–Vestfjorden (Appendix B, Figure B5a) (Olsen et al. 2001, Bergstrøm et al. 2005), Holandsfjorden (Appendix B, Figure B5b) (Olsen 2002), Trondheimsfjorden (Appendix B, Figure B5c) (Sveian and Solli 1997, Olsen et al. 2007), and Osterøyfjorden and Sørfjorden (Trengereid) northeast of Bergen (Appendix B, Figure B4) (e.g., Mangerud 1980). The ice retreat during Allerød is assumed to have reached the fjord heads and even further upstream in Møre and Romsdal County (Andersen et al. 1995) and in the Nordfjord area (Klakegg et al. 1989). The Allerød ice retreat in the Oslofjord area, based on shell dates, is not recorded to have reached more than a few kilometres proximal to the Younger Dryas (Ra) moraines (Appendix B, Figure B5d) (Sørensen 1983).

The subsequent Younger Dryas readvance was of considerable length and reached, for example, at least 10 km west of Oslofjorden (Langesund area, Figure 23), more than 40 km in the area around Bergen (Andersen et al. 1995), from a few kilometres to at least 10–20 km in the Trondheim region (Reite et al. 1999, Olsen et al. 2007, Kolstrup and Olsen 2012), and in northern Norway at least 50 km in the Ofotfjorden–Vestfjorden area (Olsen et al. 2001, Bergstrøm et al. 2005), more than 30 km in Astafjorden and some 20 km in Salangen, Troms (Vorren and Plassen 2002, Bergstrøm and Olsen 2004). Therefore, the readvance during the Younger Dryas was generally apparently more than 10 km, but one exception is known. In the fjord area west of the Svartisen glacier in northern Norway, the position of the Younger Dryas ice margin is located within a few hundred metres (to a few kilometres) from the ice margin during the Little Ice Age around AD 1723–1920 (Gjelle et al. 1995). The Younger Dryas readvance in this area must therefore have been very small, which is similar to the western part of Svalbard (Mangerud and Landvik 2007).

All these data indicate considerable regrowth of the ice, especially for the Younger Dryas with a regional readvance occurring along and filling fjords several hundred metres deep, so that the ice reached a thickness of 800–1200 m in fjords that had been ice free during the Allerød (Andersen et al 1995, Mangerud 2004). In addition, a rise in relative sea level (transgression) of up to 10 m is recorded in the same area as the late Younger Dryas readvance occurred in southwest Norway (Anundsen 1985). The interpretation has been that this relative sea-level rise was caused by the combined effect of several factors, where one of these was that the Younger Dryas ice growth was big enough to slow down or halt the general glacioisostatic uplift during deglaciation (Fjeldskaar and Kanestrøm 1980, Anundsen and Fjeldskaar 1983). At the position of the Younger Dryas ice margin near Bergen, the relative sea level was close to 60 m a.s.l. at both 13.8 and 11.4 ka (Lohne et al. 2004, Lohne 2006), with the implication that the ice load during Younger Dryas was almost the same as before the Allerød deglaciation, including the isostatic 'memory' of the LGM ice load (Mangerud 2004).

The regional differences in ice growth and ice advances may be a result of topographical and glaciological effects (Manger-

Figure 23. Map of moraines and suggested ice-retreat lines west of Oslofjorden, slightly modified from Bergstrøm (1999). Ages are in ^{14}C ka (between c. 14.1 and 11.1 cal ka, i.e., 12.2–9.7 ^{14}C ka).



ud 1980), but differences in precipitation (snow) may have been even more important. For example, southwesterly winds are supposed to have been dominant during Younger Dryas ice growth in southwest Norway (Larsen et al. 1984), which is consistent with the large Younger Dryas advance in this region.

Compilations of Lateglacial ice margins have been carried out for all parts of Norway by e.g., Andersen (1979), and for the Younger Dryas cold period more recently by Andersen et al. (1995) and Mangerud (2004). Regional ice-marginal reconstructions for the Lateglacial period that deviate slightly or even significantly from previous compilations have recently been presented for parts of Norway, e.g., the Andfjorden area (Vorren and Plassen 2002) and the Vestfjorden area, northern Norway (Ottesen et al. 2005b, Bergstrøm et al. 2005, Knies et al. 2007), Mid-Norway (Bargel 2003, Bargel et al. 2006, Olsen et al. 2007), and southwest Norway (Bergstrøm et al. 2007, Lohne et al. 2007).

It seems clear from these reconstructions that the ice margin reached at least to the middle part of the continental shelf off mid and northern Norway just before the Lateglacial period started *c.* 15.3 ka (e.g., Rokoengen and Frengstad 1999), and after *c.* 14.7 ka most ice margins had withdrawn from the shelf.

During the Younger Dryas maximum, significant ice growth and ice readvance occurred in most, but not in all western parts of the Scandinavian ice sheet. For example, based on extensive regional Quaternary geological mapping, Follestad (1994b, 2003a, b) concluded that the Trollheimen massif had only local glaciers during the Younger Dryas, and that these were dynamically separated from the Scandinavian ice sheet (Figures 24 and 25). In addition, no distinct ice-marginal moraines from the Younger Dryas maximum (Tautra) are recorded in areas above the late-/postglacial marine limit between Oppdal–Trollheimen and Stjørdal northeast of Trondheim (Reite 1990),

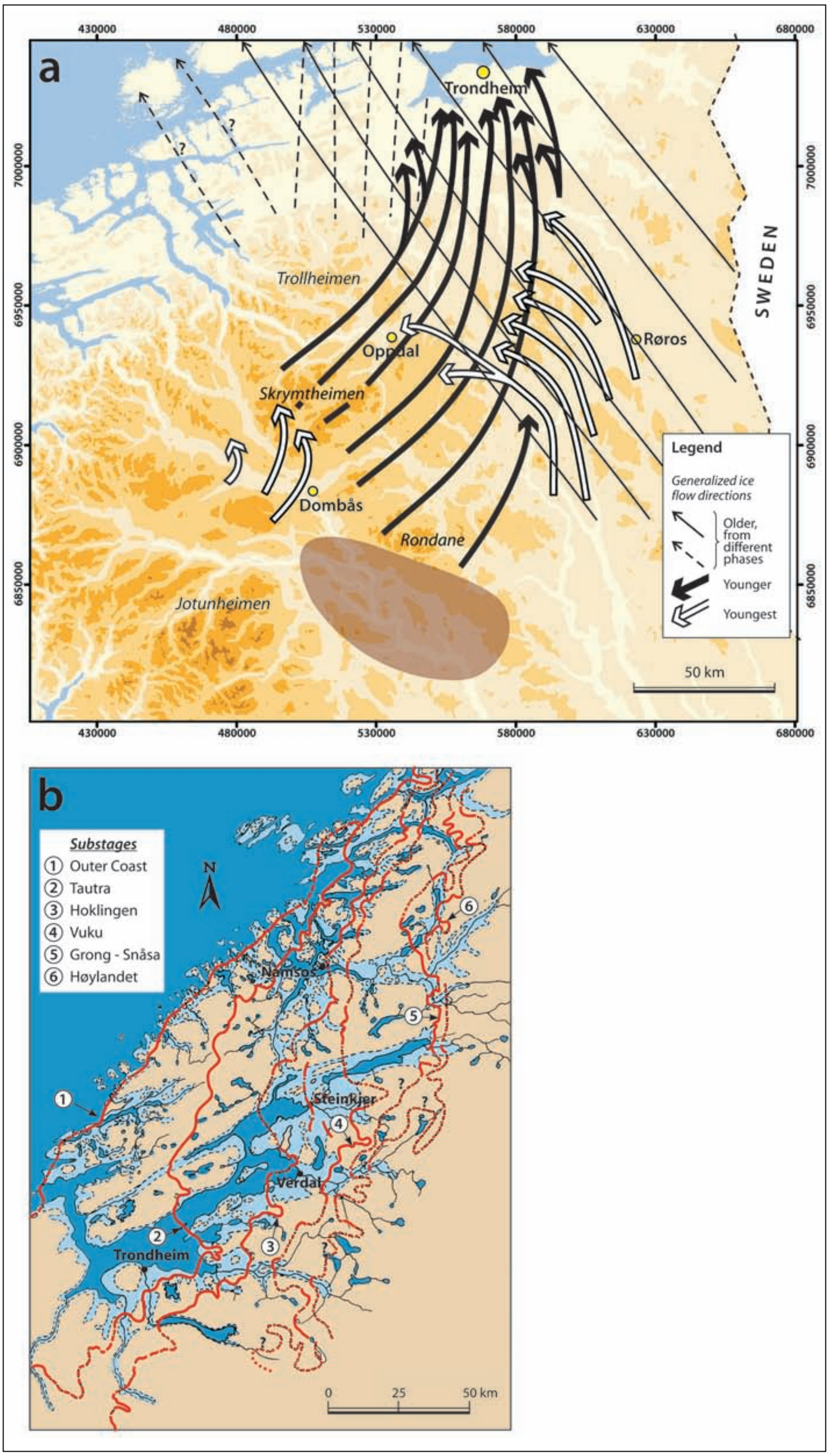


Figure 24. (a) Generalised ice-flow directions of southern central Norway during the Late Weichselian, based on glacial morphological features (modified from Follestad and Fredin 2007). The older and oldest phases represent LGM–late-glacial ice flow prior to the Allerød and Younger Dryas (Tautra Substage) (modified from Reite 1990, 1995). The younger phase may represent late Allerød and early Younger Dryas ice flows, whereas the youngest phase may correspond with the late Younger Dryas Hoklingen Substage (see b). The shaded area in the south indicates a zone of divergent N-S ice movements (ice divide zone) during the LGM interval. (b) Lateglacial ice margins in the Trondheimsfjord area (modified from Sveian 1997).

which indicate an insignificant early Younger Dryas readvance in this area. Further mapping and deglaciation studies in the areas southeast of Trollheimen have recently led to a new understanding of the Younger Dryas horizontal as well as vertical ice limit in these central areas of glaciation. Lateral moraines of assumed early Younger Dryas age and representing the Scandinavian ice sheet are situated about 1200 m a.s.l. at the eastern flank of Trollheimen (Follestad 2003a, b, 2005a, b), with locations more than 100 km north of the inferred ice divide. Comparison with possible modern analogues from Eastern Antarctica (Kapitsa et al. 1996, Bell et al. 2007) suggest that the ice surface during Younger Dryas maximum, at the

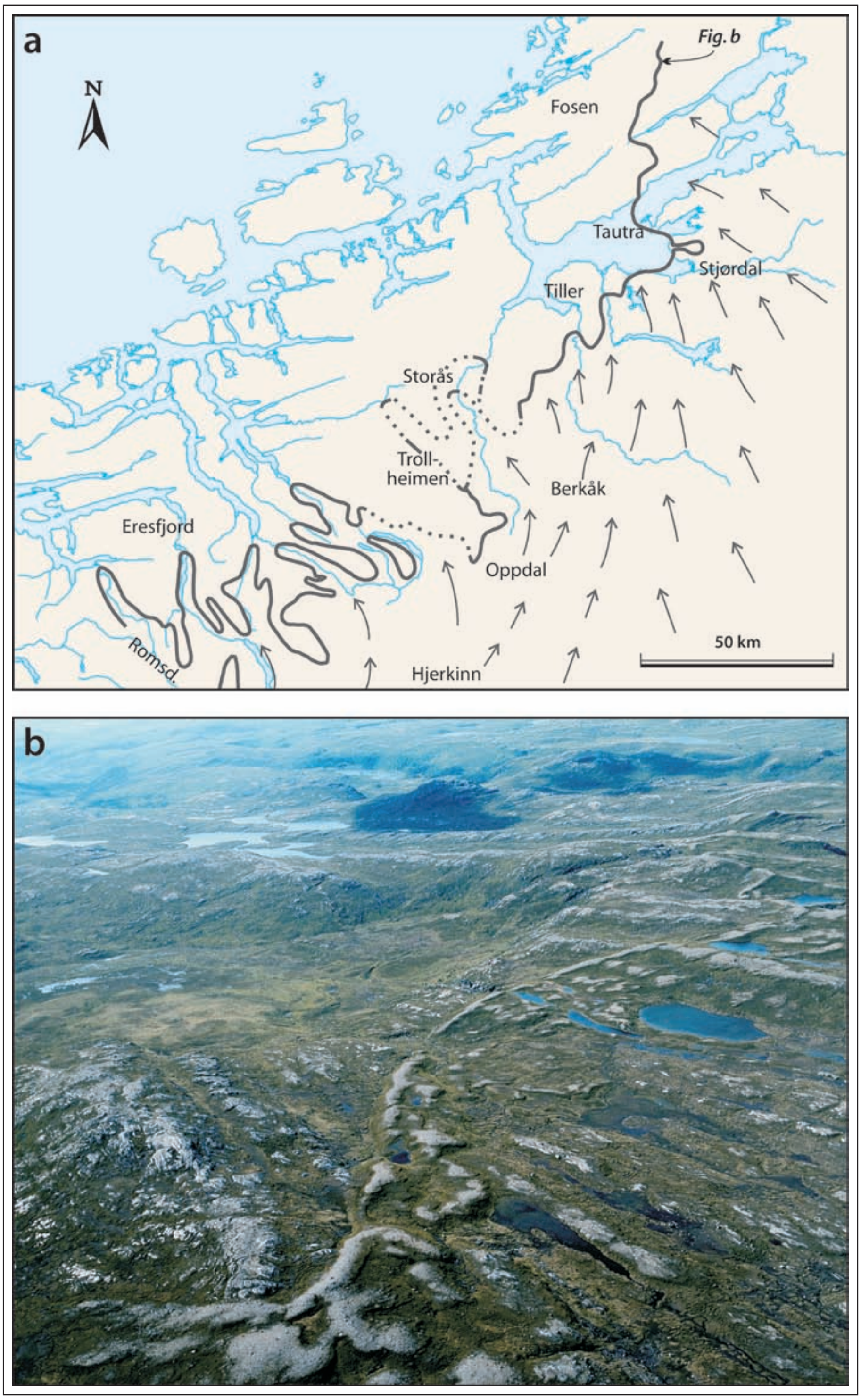


Figure 25. (a) Map of the reconstructed Younger Dryas ice margin represented by the Younger Dryas (Tautra) moraines from Trondheimsfjorden in the north to Romsdalen–Hjerkinn in the south, modified from Follestad (2005b). The stippled lines around Trollheimen indicate more uncertainty in the correlations there than elsewhere in the area. (b) The Tautra moraine at Fosen. Photograph by H. Sveian.

ice divide east of Vinstra in Gudbrandsdalen, may have been as low as 1400–1500 m a.s.l. if much water occurred in big subglacial basins in the zone between the ice divide and the zone of lateral moraines (Trollheimen/Oppdal/Sandfjellet), and at least to 1500–1600 m a.s.l. if such basins did not exist (Figure 26, Olsen et al. 2007). Most recently, Follestad (2007) has mapped lateral Younger Dryas moraines up to *c.* 1400–1500 m a.s.l. south of Lesjaskog, west of northern Gudbrandsdalen, that match well with the theoretical Younger Dryas ice-surface profile in Figure 26. The late Younger Dryas Hoklingen lateral moraines have recently been mapped up to *c.* 1100 m a.s.l. at Sandfjellet (Olsen et al. 2007) some 30 km north of Kvikne.

This constrains the late Younger Dryas ice surface in that area, which is located around the water divide (Kvikne, 700 m a.s.l.) some 100–130 km north of the ice–divide zone (Figure 26).

One problem that hamper the reconstruction of ice limits, ice flows and accompanying melt-water drainage patterns in central southeast Norway is the apparent concentration of multistage glacial features, e.g., different sets at the same location of crossing lateral meltwater channels, that are found here (Sollid and Sørbel 1994, Follestad and Thoresen 1999, Follestad 2005c, 2006, Olsen et al. 2013, fig. 9). Which features belong to the youngest phase and which are older? This problem has to be solved before a reliable deglaciation model can be established. Multistage glacial features are well known from areas close to the ice divide also in other parts of Scandinavia (Lagerbäck 1988, Kleman 1990, 1992, 1994, Kleman and Borgström 1990, 1996, Kleman et al. 1997, Hättestrand 1998, Hättestrand and Stroeven 2002), but less frequently occurring in areas farther away from the ice divide (e.g., Olsen et al. 1987, Riiber et al. 1991). The only reasonable glaciological explanation for the existence of such multistage features is to include phases of cold-based ice that have preserved the glacier bed from later erosion. The understanding of the dynamic development both in time and space of the different glacial temperature regimes and their effects on the landscape evolution in central southeast Norway are at present merely in an initial phase. However, much regional Quaternary geological mapping has recently been carried out here (Follestad and Thoresen 1999, Follestad 2003a, b, 2005a, b, c, 2006, 2007, Quaternary map of Hedmark county, in prep.), and we predict that a quite new understanding of the deglaciation in these areas, based on new well-constrained field data will soon be achieved and presented.

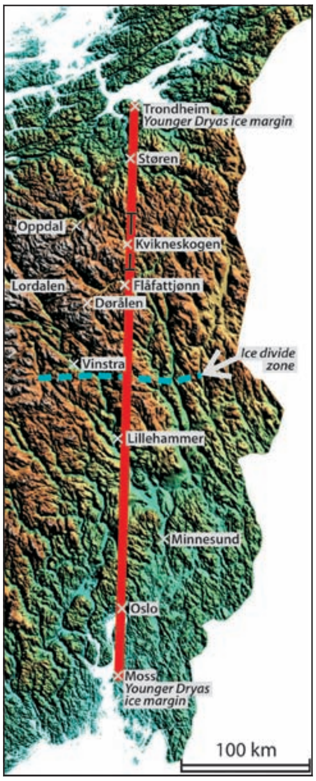


Figure 26. Ice-surface profiles for the early Younger Dryas (Tautra) and late Younger Dryas (Hoklingen) substages from Trondheim to Moss are outlined. Location of site areas and transect of profiles are shown to the left. After Olsen et al. (2007). Scenarios with and without glacial isostasy are indicated. Glacioisostatic depression is based on data from Eriksson and Henkel (1994), Eronen (2005) and Påsse and Andersson (2005). Younger Dryas lateral moraines located up to 1400–1500 m a.s.l. and around 1200 m a.s.l., are projected into the profile from Lordalen (Lesjaskog), Oppdal and eastern Trollheimen (after Follestad 2003a, b, 2005a, b, 2007). X–Deglaciated down to this level prior to Younger Dryas according to Paus et al. (2006). Y–Deglaciated down to 1000–1100 m a.s.l. during late Younger Dryas according to Dahl et al. (2005).

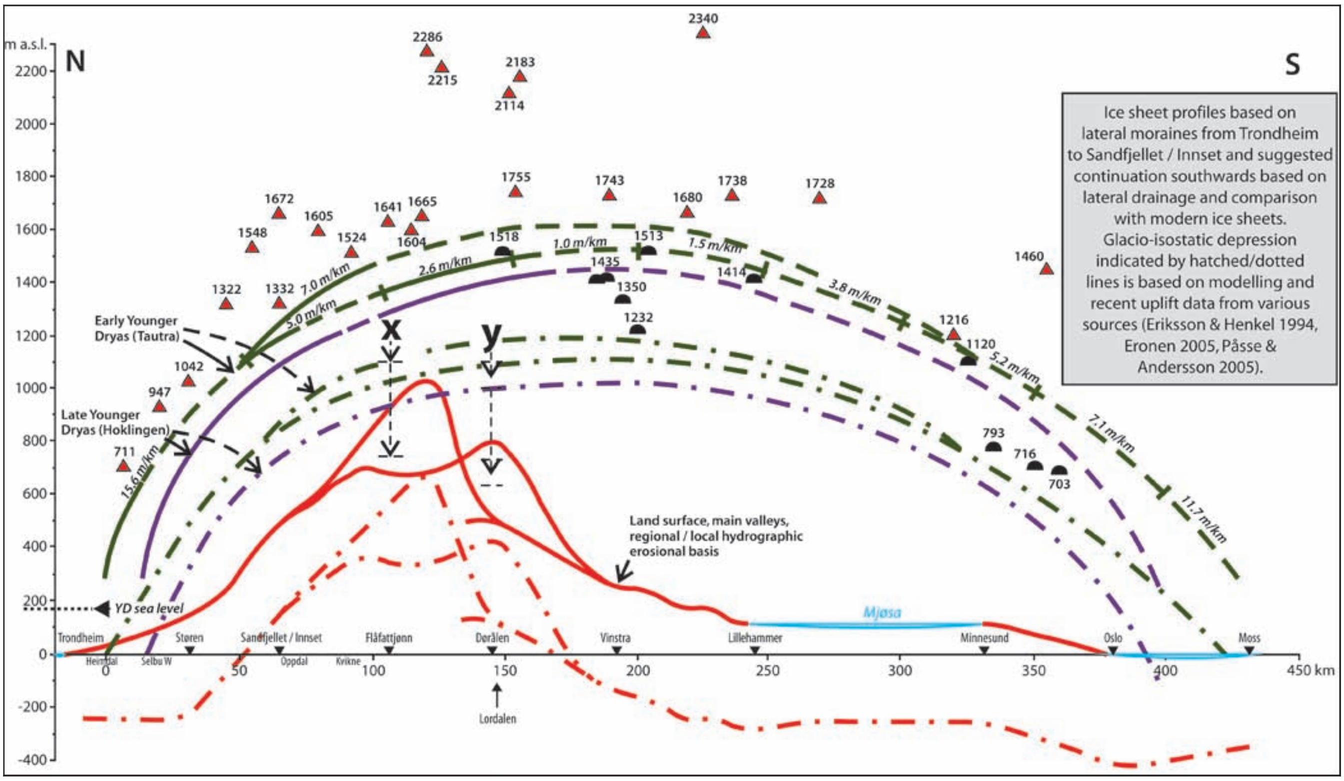


Figure 27. The distinct late Preboral Ølfjell Moraine at Saltfjellet, North Norway (Sveian et al. 1979). View from Ølfjellet towards the north. Photograph by S. Gjelle.

Lateglacial/Early Holocene ice-dammed lakes in central southeast Norway

The long debate on the occurrence and character (open or partly ice-filled lake) of the Preboreal glacial lake 'Nedre Glåmsjø', southeast Norway (Holmsen 1915, 1960, Gjessing 1960, Andersen 1969, Østeraas 1977, Sollid and Carlson 1980) was apparently terminated with the studies of Longva and Bakkejord (1990), Longva and Thoresen (1991), and Longva (1994). Longva (1994) concluded that the flood event, which resulted from the catastrophic drainage of Nedre Glåmsjø was of a size that indicated a water body similar to that of an open Nedre Glåmsjø lake in its whole length. However, recent mapping by NGU in this area, as mentioned above, has provided data that challenge the open-lake theory and shed new light on the old ideas of an overall subglacial drainage during deglaciation, presented by Gjessing (1960). Subglacial drainage from higher elevations is included directly in the ice-dammed lake environments. It is concluded that most of the waterlain deposits within the entire ice-dammed lake area, including older more extensive and younger less extensive parts, have been deposited subglacially in tunnels and chambers of various sizes (e.g., Follestad 1997, Olsen and Sveian 2005). The present authors do not dispute the water volume needed in Longva's (1994) flood model, but we suggest that the catastrophic drainage occurred at an earlier phase during the last deglaciation with more ice present in the valley systems, so that the sum of local lateral/sublateral ice-dammed lakes and many water-filled subglacial chambers and tunnels would add up to a size comparable to that of an open Nedre Glåmsjø lake.

The flood event occurred *c.* 9.1–9.2 ^{14}C ka (10.24–10.33 cal ka) (Longva 1994) with the ice surface in the Nedre Glåmsjø area reaching at least 670–700 m a.s.l. (Longva 1994, Olsen and Sveian 2005). A more precise elevation of the ice surface is difficult to estimate, but a maximum must be well below 1100 m a.s.l. since the vertical lowering of the ice surface reached this level in the western part of the lake area before 9.5–9.6 ^{14}C ka (10.7–10.9 cal ka) (Paus et al. 2006). A revision of the model for the catastrophic drainage of the Nedre Glåmsjø ice-dammed lake, as we suggest here, implies a minor reduction in the melting rate of the ice thickness, to less than 400 m (<1 m yr^{-1}) between 9.6 and 9.1 ^{14}C ka (10.9–10.2 cal ka).

In core MD99–2286 from the Skagerrak Sea outside the Oslofjord, an IRD peak at 9.1–9.2 ^{14}C ka (10.4–10.5 cal ka) was correlated to the Nedre Glåmsjø flooding event by Gyllencreutz (2005). About 500 years earlier, the last calving ice margin receded onshore in the Oslofjord area at *c.* 9.7 ^{14}C ka (11.1 cal ka), based on mapping of the marine limit (Hafsten 1983) and dating of the ice-marginal ridges both above and below the marine limit (Andersen et al. 1995).

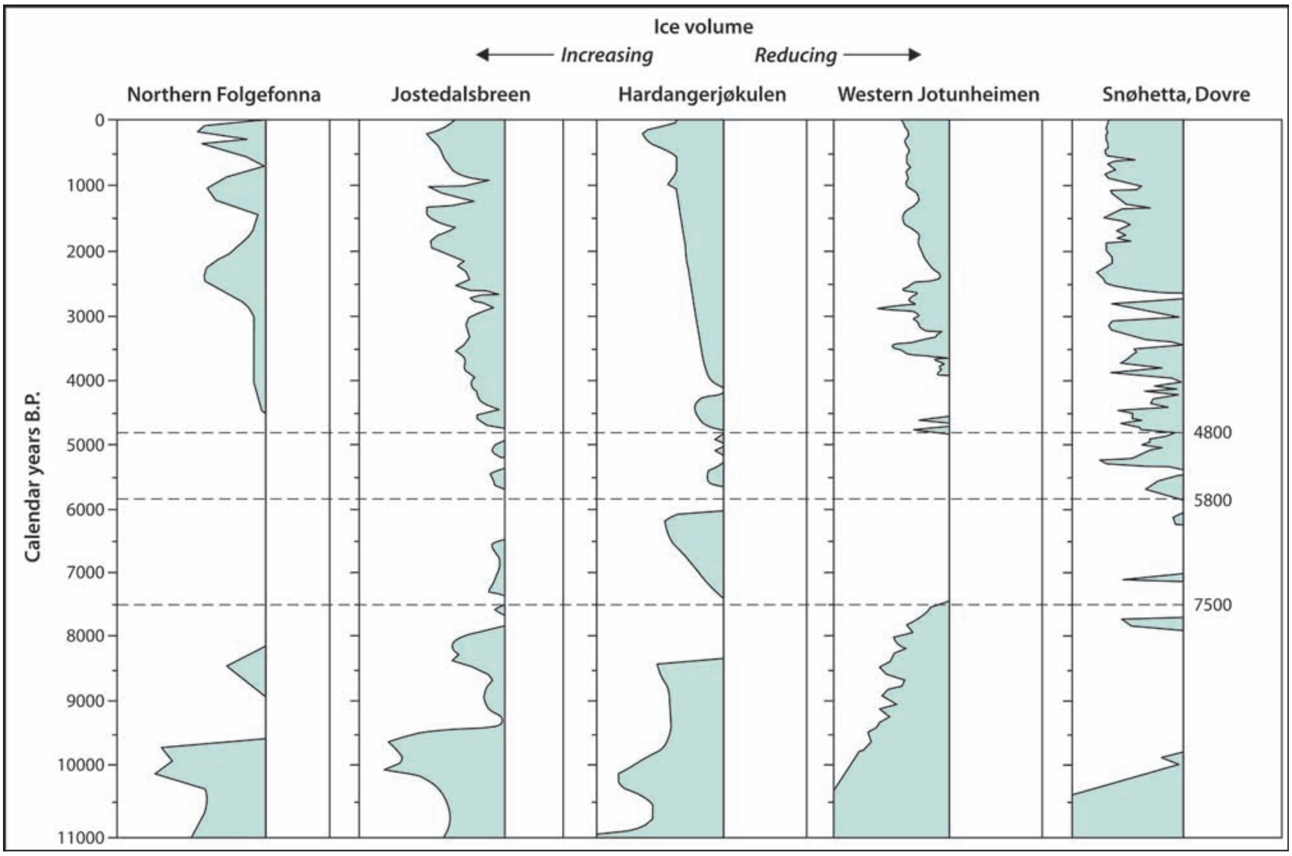


Figure 28. Variations in ice volume for five glaciers in southern Norway during the Holocene. Slightly modified from Larsen et al. (2003). All glaciers seem to have been totally deglaciated (almost no ice left) in short periods around 7500, 5800 and 4800 calendar years before present (B.P.) (stippled lines).

Glacial variations during the Holocene

The Holocene is mainly dealt with by Fredin et al. (this issue), but the glacial variations in this period are (also) briefly mentioned below.

Ice-marginal moraines representing one or more glacial events at the transition between Younger Dryas and Preboreal are reported from most parts of Norway, but most of these seem to represent insignificant readvances, with the Nordli/Vuku Substage in Nordland and Trøndelag as a distinct exception. This glacial event includes apparently a considerable local ice readvance (Andersen et al. 1995, Sveian and Solli 1997). Glacial readvances or merely halts during ice retreat are known from regional mapping of ice-marginal deposits also for the time around 9.7 ± 0.2 ^{14}C ka (11.1 cal ka) and 9.3 ± 0.2 ^{14}C ka (10.6 cal ka), for example in Troms and Nordland (Andersen 1968) (Figure 27, Ølfjellmorenen, Sveian et al. 1979, Andersen et al. 1995, Blake and Olsen 1999), in inner Sogn (Vorren 1973, Bergstrøm 1975), in southwest Norway (Blystad and Selsing 1988, Bang-Andersen 2003, Bergstrøm et al. 2007), and in southeast Norway (Sørensen 1983, Olsen 1985a).

Holocene glacier fluctuations are traditionally recorded from the distribution of moraines (e.g., Andersen and Sollid 1971), but also from studies of lacustrine and glaciolacustrine sediments in basins close to present glaciers, and several glacial events are recorded both in southern Norway (Figure 28, and Appendix B, Figure B1) (Matthews et al. 2000, Nesje et al. 2001, Bakke et al. 2000, 2005a, Nesje et al. 2008) and northern Norway (Bakke et al. 2005b). For example, distinct ice-growth events are recorded around 9.3–9.1 ^{14}C ka (10.6–10.4 cal ka) (Erdalen event, Dahl et al. 2002, 'Lønsdal event', Sveian et al. 1979), *c.* 7.2 ^{14}C ka (corresponding to the Finse event, *c.* 8.2 cal ka, Dahl and Nesje 1996, Nesje et al. 2001), *c.* 6 ^{14}C ka (6.85 cal ka), *c.* 4.0–2.5 ^{14}C ka (4.44–2.7 cal ka), and during the 'Little Ice Age' *c.* AD 1500–1920, with culmination *c.* AD 1750, 1780–1820, 1830, 1850, 1870–1890, 1910 and 1930 at different glaciers existing in continental to maritime climate regimes (e.g., Bakke et al. 2005b).

The Little Ice Age event has generally included the most extensive glacier advances during the middle to late Holocene. Glacier advances beyond the Little Ice Age glacier maximum in this age interval have so far only been reported from *c.* 3–2 ^{14}C ka (3.2–1.95 cal ka) at Okstindan (Griffey and Worsley 1978) and at Folgefonna (Bakke et al. 2005a).

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Appendix A

Glacier extension maps for three pre-Weichselian stages

The glacier extension maps (Figure A1) show the maximum glacier extensions areally, with the assumed position of the ice margin indicated. The data background for northern and central Fennoscandia is the same as used in the glaciation curve (Figure 2, the main text). The time-transgressive pre-Weichseian ice-margin positions during maximum glacier extensions are indicated for comparison on each map. This line corresponds in several areas approximately to the ice extension during the Saalian (sensu stricto), in other parts to the Elsterian (MIS 12) and possibly to the Don glaciation (at least 500 ka, possibly 650 ka) in the southeast.

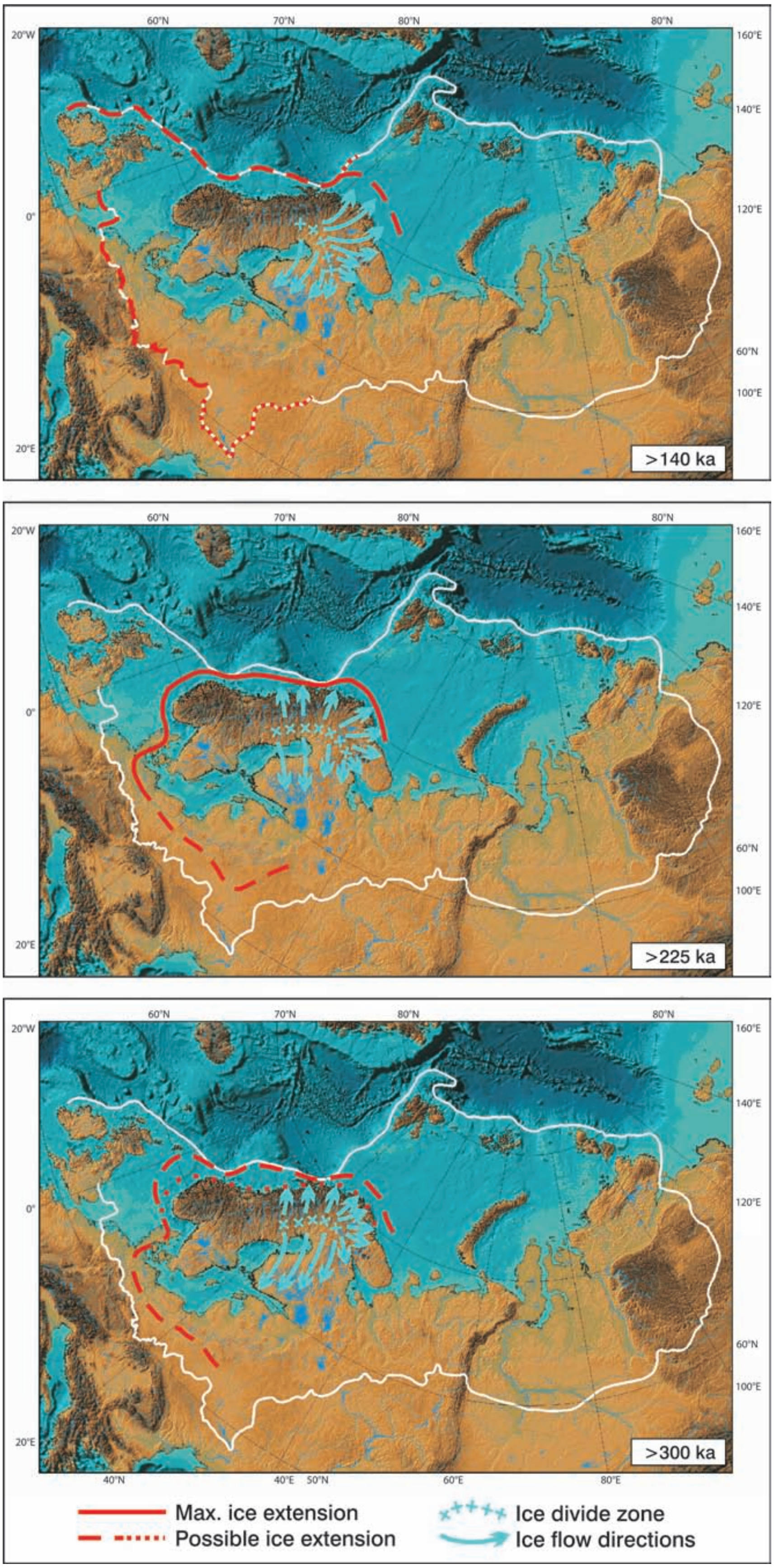

Figure A1. Glacier extension during three pre-Weichelian stages, thought to correspond to MIS 6 (upper panel), MIS 8 (middle panel) and MIS 10 (lower panel), based mainly on stratigraphical data from northern and central Fennoscandia. After Larsen et al. (2005). The correlation with data from the adjacent shelves and from the mainland in the south and southeast are based on Dahlgren (2002), Dahlgren et al. (2002), Nygård (2003), Rise et al. (2005, 2006) and Ehlers (1996).

Appendix B

Additional glaciation curves

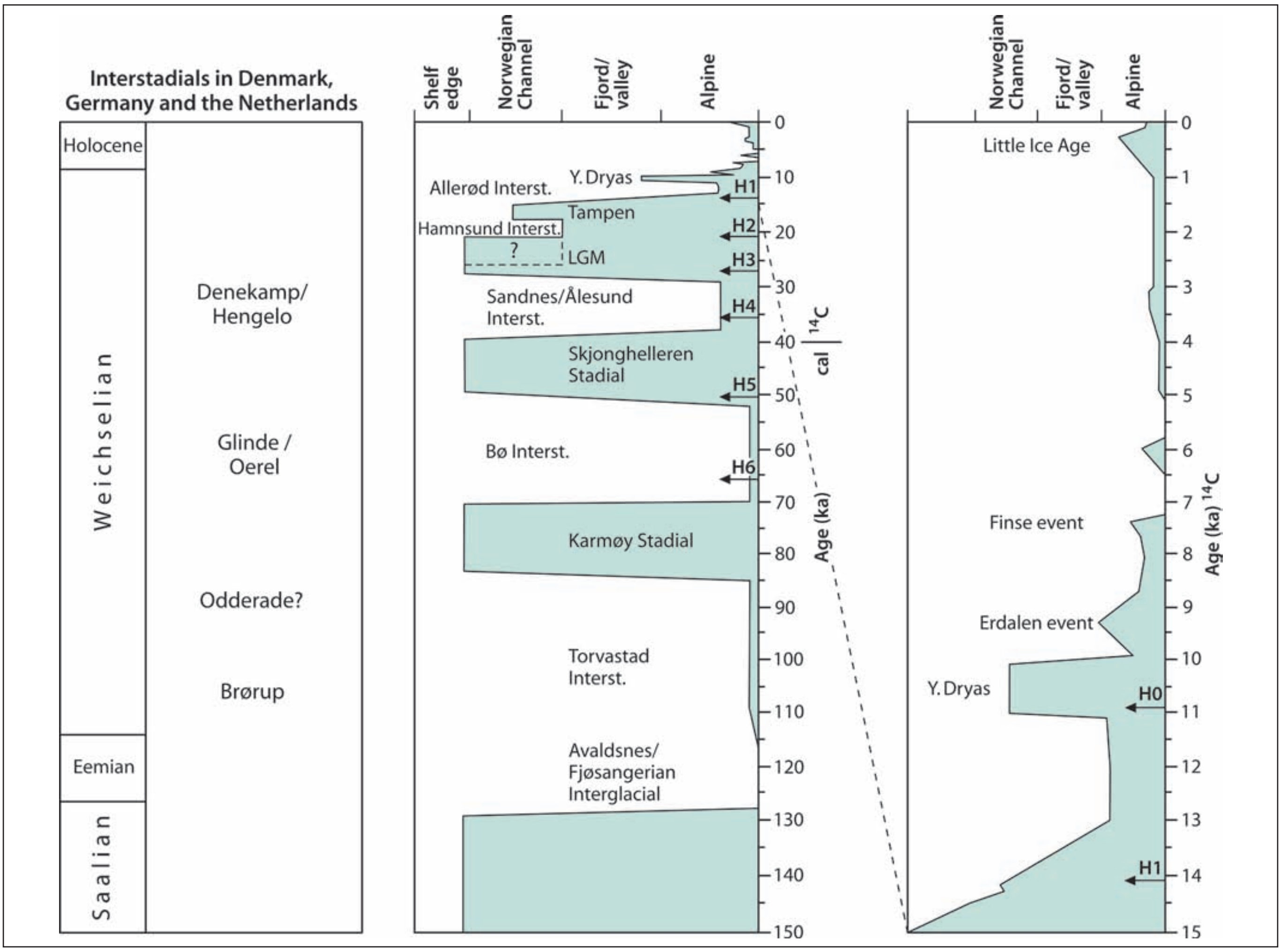

Figure B1. Glacial curves for the last 150,000 years from southwest Norway and adjacent sea-bed areas. H0, H1, H2, H3, H4, H5 and H6 indicate suggested positions of Heinrich events H0–H6. Modified from Sejrup et al. (2000). Note that the age scales are in ^{14}C ka at ages <40–45 ka and cal ka for older ages.

Comparison between ice growth in horizontal and vertical direction of the Fennoscandian/Scandinavian ice sheet

A reconstruction of relative size variations, both in area and ice volume for the Fennoscandian/Scandinavian ice sheet during the Weichselian glaciation was presented recently (Olsen 2006) (Figure B2). These curves are based on Figure 9 (in the main text) for horizontal ice extension, and from core data from the Vøring Plateau presented by Baumann et al. (1995) for the estimated ice volumes. The correlations between the curves should be considered as tentative since the age models used for the terrestrial and deep-sea sediments are different (based mainly on radiocarbon and luminescence dates for the terrestrial samples, and radiocarbon dates and oxygen isotope data for the marine sediments), and several of the 'ups' and 'downs' in the curves are therefore wigglematched. However, despite these flaws the curves indicate relative values closer to each other during glacial advances (maxima) than during retreats (minima). This trend is more pronounced during younger ages and seems to correspond to the increasing sizes of the advances towards younger ages. The generally larger difference in relative sizes for ice extension and ice volume during ice retreats is quite as expected from a topographical point of view, particularly in the west (Norway) with the many fjords and therefore the good calving conditions during ice retreat. In such situations the horizontal ice retreat would be expected to go much faster than the general reduction in ice thickness (ice volume). Also the ice growth during almost or fully maximum ice-sheet conditions should not be expected to match fully in horizontal and vertical direction. This is shown from modelled glacier growth for the last part of the LGM in Fennoscandia (Charbit et al. 2002). It is also shown from Greenland and Antarctica (Huybrechts 2002), where maximum ice volume during LGM lags maximum ice extension by several thousand years (4,500–10,000 years).

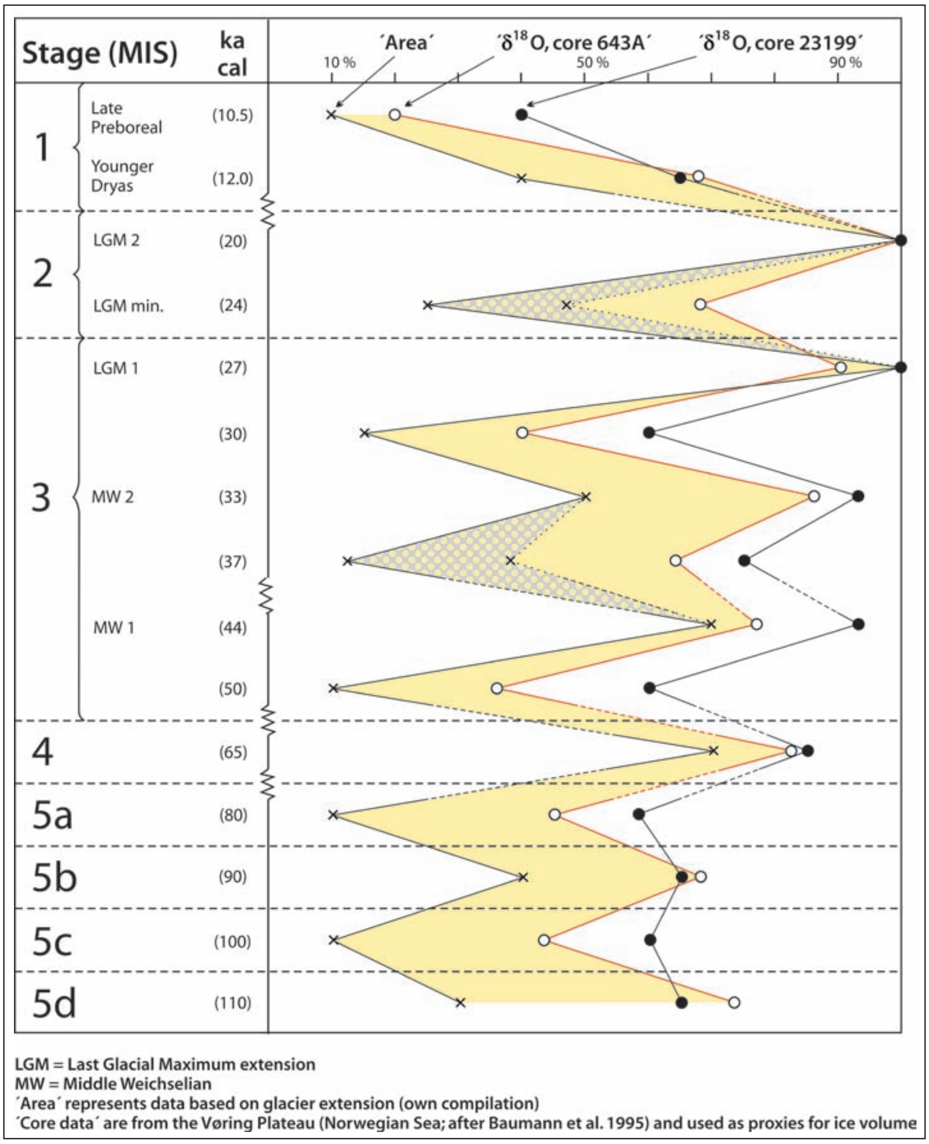

Figure B2. Relative size variations of the Fennoscandian/Scandinavian ice sheet, 110–10 ka. Comparison between horizontal ice extension and ice volume. The area glacier variations are in accordance with Figure 9 (main text), and the ice-volume data are from the Norwegian Sea (Baumann et al. 1995). After Olsen (2006). Note the changes in time scale at four places, between MIS 1 and 2, 3 and 4, 4 and 5a, and within MIS 3.

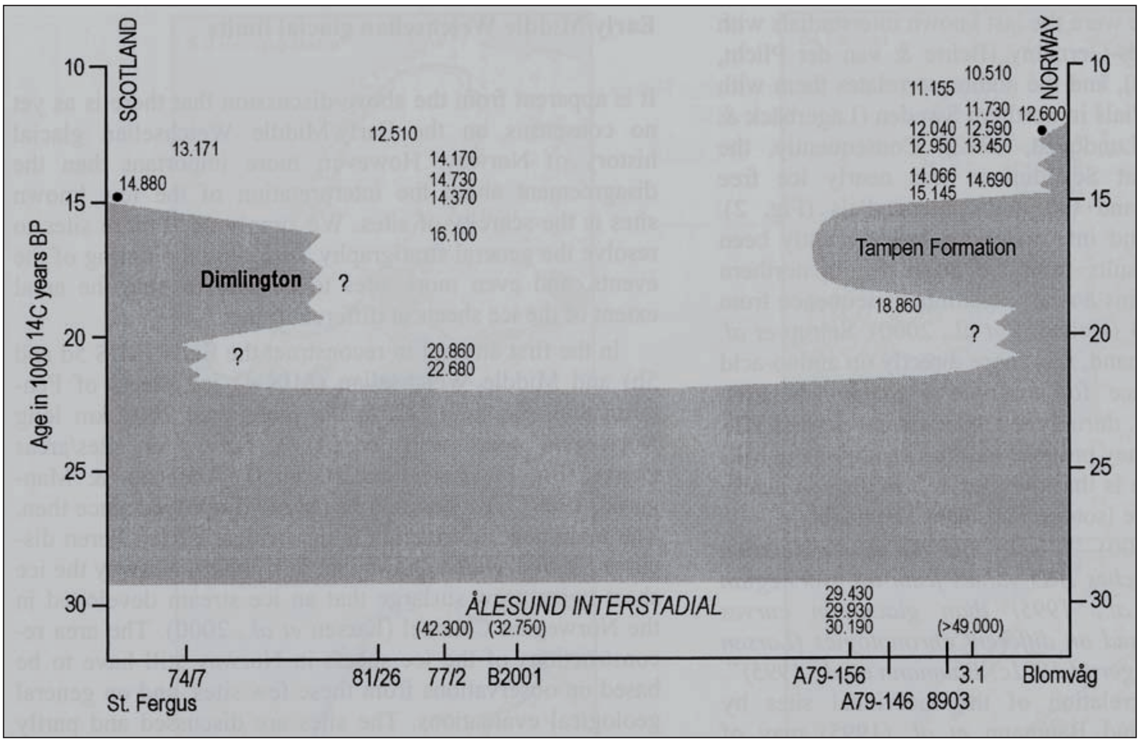

Figure B3. Glacial (time–distance) curve for the British and Scandinavian ice sheets across the North Sea during the LGM interval. The two ice sheets are supposed to have met before c. 22.7 ^{14}C ka (c. 26.7 cal ka). Ages are in ^{14}C yr BP. After Sejrup et al. (2000).

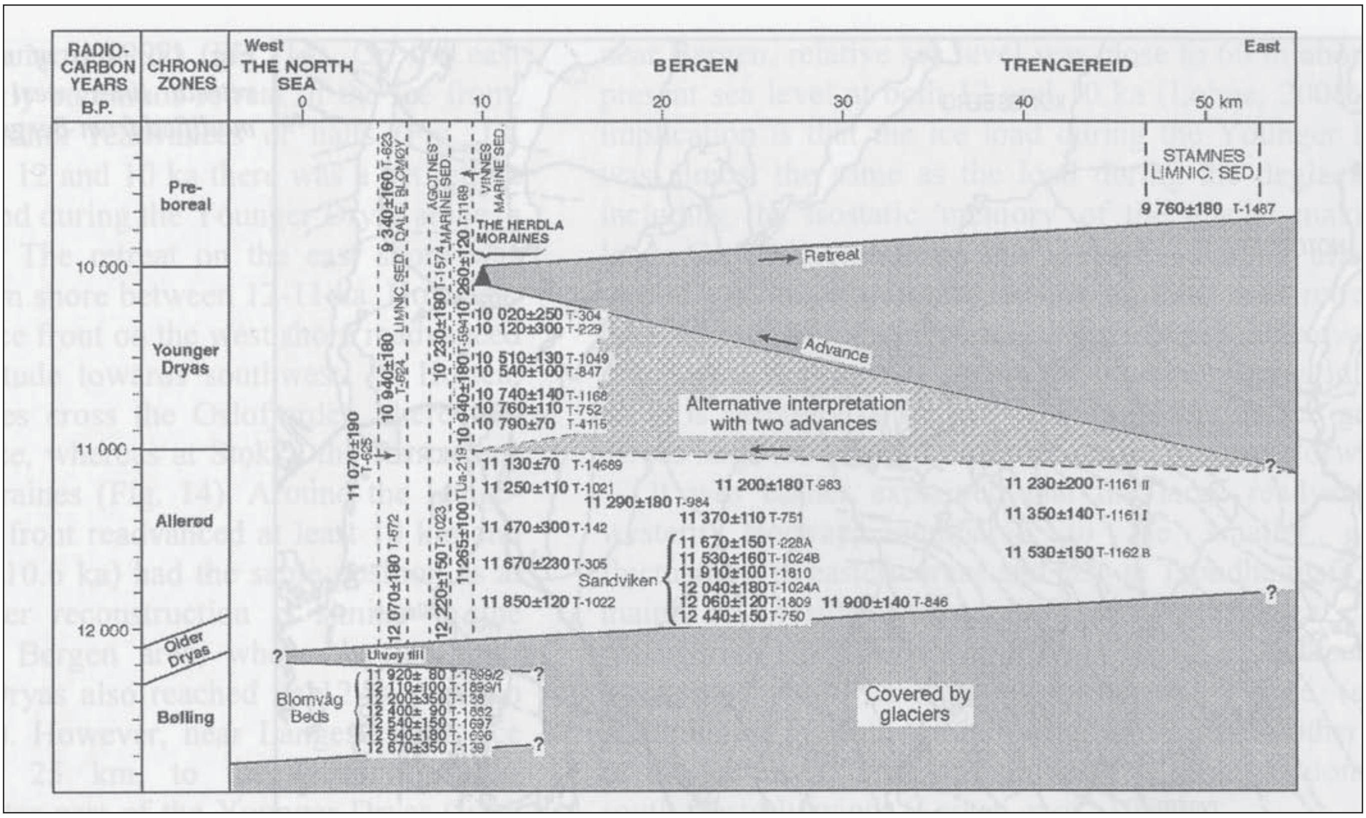

Figure B4. Ice-front fluctuations in the Bergen district, from Mangerud (2000). Ages are in ^{14}C yr BP. A generalised version of this diagram is included as diagram d in Figure B5.

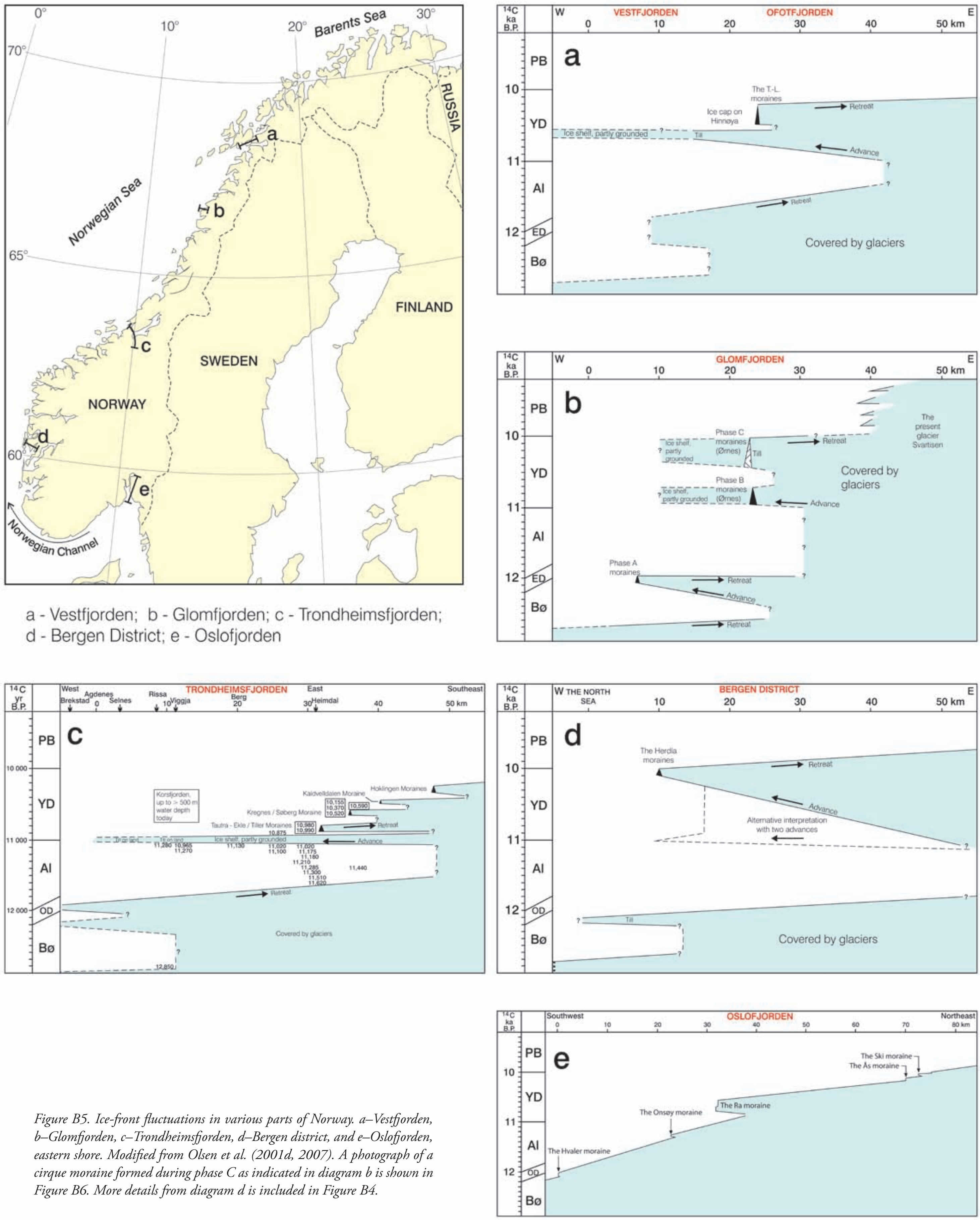

Figure B5. Ice-front fluctuations in various parts of Norway. a–Vestfjorden, b–Glomfjorden, c–Trondheimsfjorden, d–Bergen district, and e–Oslofjorden, eastern shore. Modified from Olsen et al. (2001d, 2007). A photograph of a cirque moraine formed during phase C as indicated in diagram b is shown in Figure B6. More details from diagram d is included in Figure B4.

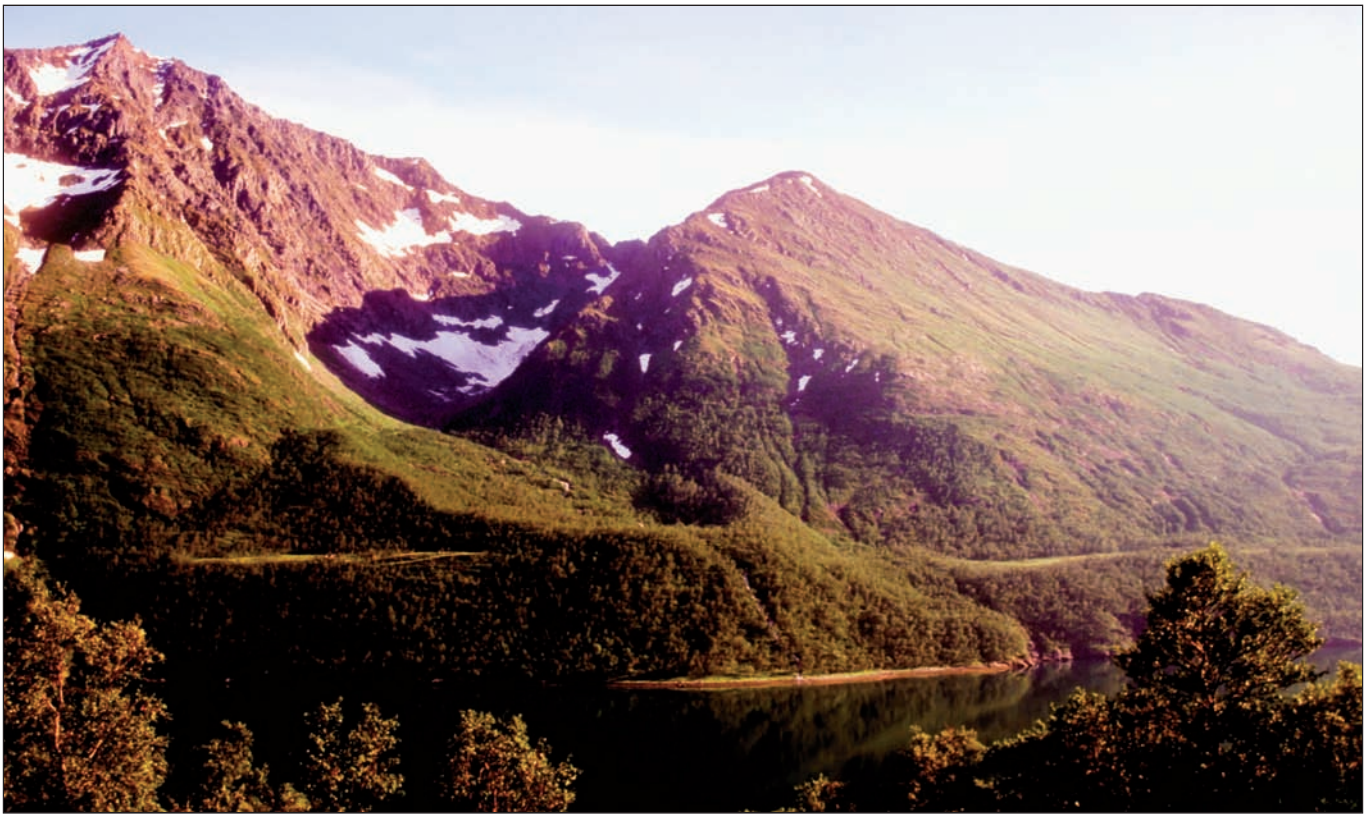

Figure B6. Cirque (-glacier) moraine at the southern side of Holandsfjorden west of Svartisen in Nordland (see Figure 3, the main text). The cirque moraine from late Younger Dryas (during phase C in Figure B5, diagram b) is located partly on a shoreline (c. 100 m a.s.l.), which was eroded in bedrock during early Younger Dryas (during phase B in figure B5, diagram b). The inland ice margin during this period was located at the heads of the fjords in this region. After Rasmussen (1981) and Olsen (2002). CN-dates from the top of the mountain to the left (max. height c. 1452 m a.s.l.), which is located only 200-300 m higher than the adjacent present glacier Svartisen, indicate almost no surficial erosion there during the last glaciation (after Linge et al. 2007).

Quaternary glacial, interglacial and interstadial deposits of Norway and adjacent onshore and offshore areas

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The late Pliocene and Pleistocene sediment distribution, thickness and stratigraphy of Norway and the adjacent continental shelf are reviewed in this paper.
A generalised map of Quaternary[1] sediments in Norway in scale 1:1 million or larger (available from www.ngu.no), compiled from Quaternary maps in various scales, and a simplified map with distribution of an overburden of >1 m-thick loose deposits are used for this regional overview. In addition, the sediment distribution and thickness of the Quaternary sediments from selected parts of the adjacent shelf areas are presented, and the geological implications (with erosion and deposition) of all these data are discussed.

The Quaternary stratigraphy from the land areas in Norway is also presented, with examples from northern, central and southern parts of the country. Correlations with adjacent areas, mainly from central and northern Fennoscandia, are suggested, and the regional implications are indicated, particularly for the Weichselian glaciation, and further dealt with in the accompanying paper focused on Quaternary glaciations (this issue).

It is now known that a major part of the present, remaining, onshore glacial deposits in Norway derive from the last glaciation (Weichselian) and that more than 90% (>100,000–150,000 km^3) of the Quaternary glacial erosional products from Norway have been transported and deposited offshore. These sediments are mainly deposited in large depocentres at the Mid-Norwegian shelf and in trough-mouth fans at the mouths of the Norwegian Channel and the Bjørnøyrenna Trench.

Olsen, L., Sveian, H., Ottesen, D. and Rise, L. (2013) Quaternary glacial, interglacial and interstadial deposits of Norway and adjacent onshore and offshore areas. *In* Olsen, L., Fredin, O. and Olesen, O. (eds.) *Quaternary Geology of Norway*, Geological Survey of Norway Special Publication, **13**, pp. 79–144.

[1] Quaternary is here used as described by Olsen et al. (this issue).

Introduction

A general map of Quaternary sediments in Norway in scale 1:1 million or larger (available as 'best of' version on the internet: http://www.ngu.no/kart/losmasse/) and a simplified map of an overburden with thick sediments are compiled from Quaternary geological maps in various scales (Figure 1). In addition, the sediment distribution and thickness of the Quaternary sediments on the adjacent shelf areas (Figures 2 and 3a, b, c) and the Quaternary stratigraphy from the land areas are presented (extended background data are included in Appendix A), and the geological implications of these data are considered and used in reconstructions shown on maps and in glacial curves (i.e., time–distance diagrams). The pattern of Quaternary ice-sheet erosion and deposition is recently considered theoretically for all of Fennoscandia (Kleman et al. 2007). This is acknowledged by the present authors and the background data for this pattern in Norway is considered and documented with some examples here. The geographical position of most places and sites, which are referred to in the text is shown in Figure 4, with more locations and names presented in Figure B1 and Table B1 (Appendix B).

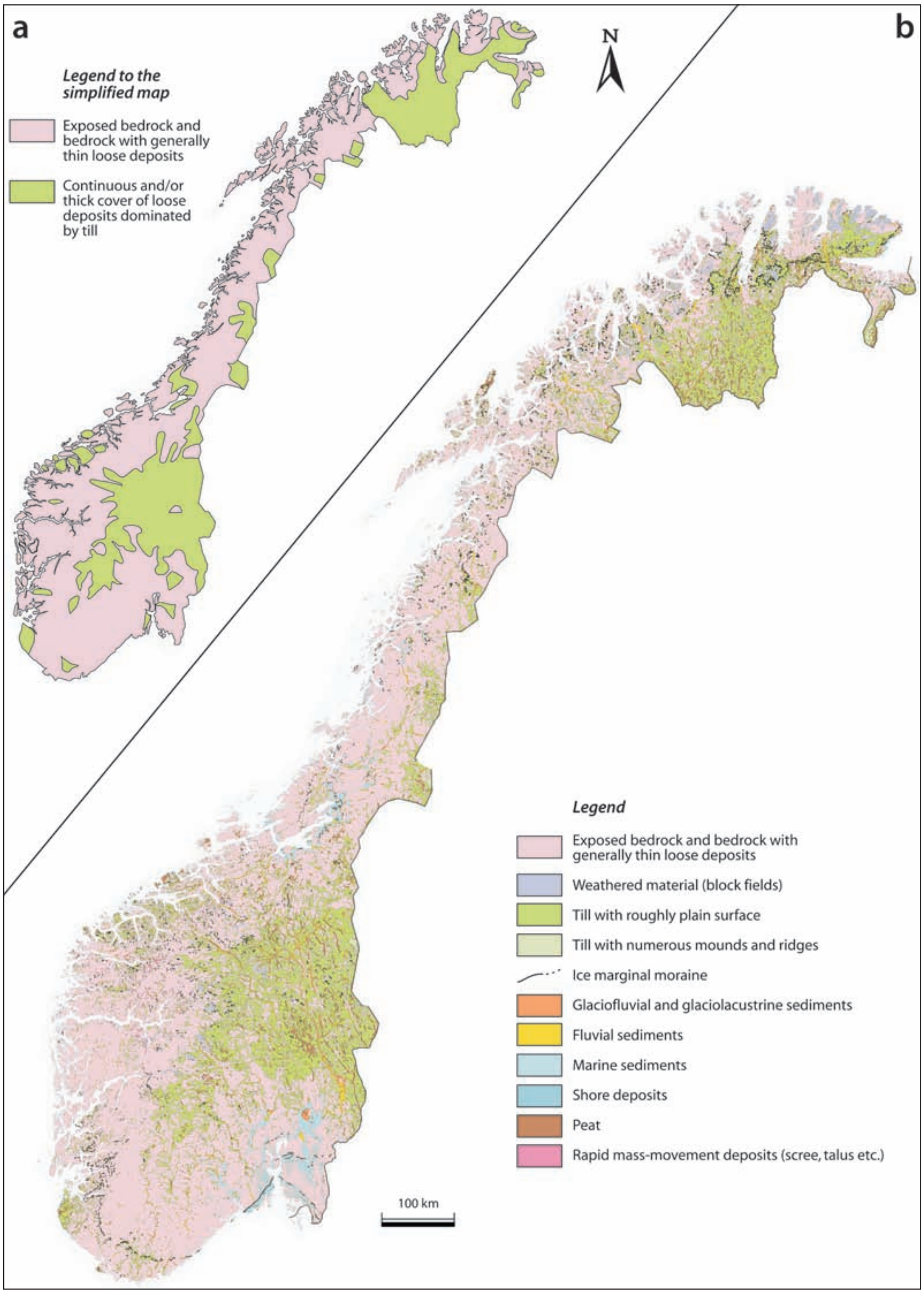


Figure 1. Quaternary geological maps of the mainland of Norway (between c. 58°-71°N and c. 5°-30°E), based on data from the NGU map database, Quaternary map of Norway. (a) Simplified map with exposed bedrock and overburden generally >1 m thickness; and (b) distribution of various sediments from Quaternary map in scale 1:1 million. The legend is simplified from the 1990-version of the map (Thoresen 1990). Some of the areas with exposed bedrock (pink) *may have a thin cover of loose deposits, but in most of these areas the overburden is assumed to be less than 1 m thick. The average thickness of the till deposits* (green)*, which cover about 25% of the mainland of Norway is estimated to* c. *6 m. For more map information, see www.ngu.no, and for further information of the sediment thicknesses, see the main text.*

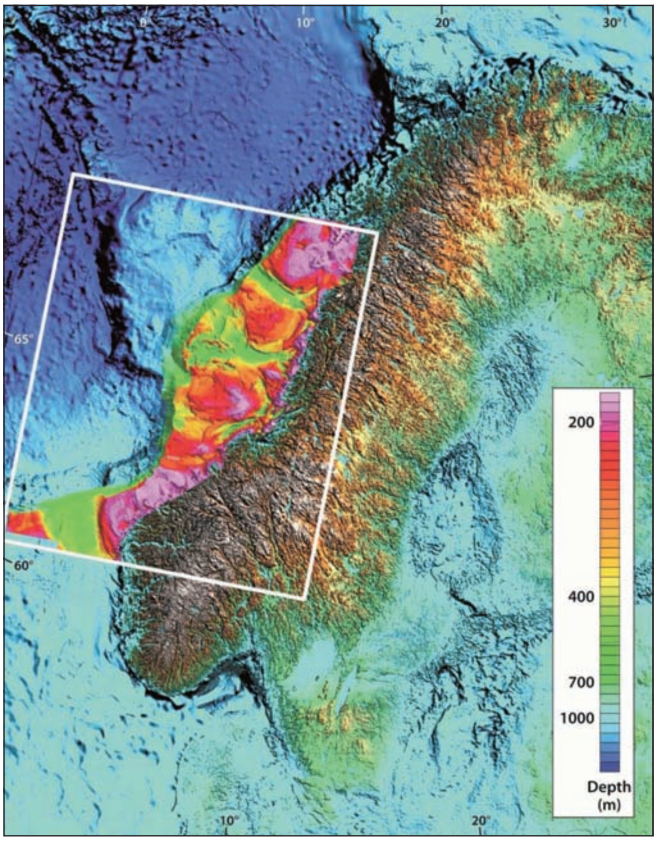


Figure 2. Topographic/bathymetric map of Norway and adjacent land and seabed areas. Position of Figure 3b is indicated. For references, see the main text.

^{14}C ages are calibrated or converted to calendar years according to Olsen et al. (2013), and the abbreviation ka BP, meaning a thousand years before present, is generally referred to as ka throughout this paper.

Shelf areas

Regional mapping programs of the Quaternary sediments on the North Sea and Norwegian shelf areas have taken place during the 1970–1980s (e.g., Rise et al. 1984, King et al. 1987).

In the last few years, during several joint industrial projects, extensive studies of the Late Pliocene/Pleistocene development (Naust Formation) of the Norwegian shelf and margin between 61°N and 68°N have been carried out. In the Ormen Lange project, the shelf and margin stratigraphy were studied by Rise et al. (2002, 2005a) on the basis of a large seismic data base. These data have given a new understanding of the environment/processes during deposition of the Naust Formation, especially the older parts. The Naust Formation has been given a unified stratigraphy across the whole area, and is subdivided into five sequences (N, A, U, S, T from bottom to top) (Figure 3c, d), each comprising several units (Rise et al. 2006, Ottesen et al. 2007). The mapping of these five sequences has made it possible

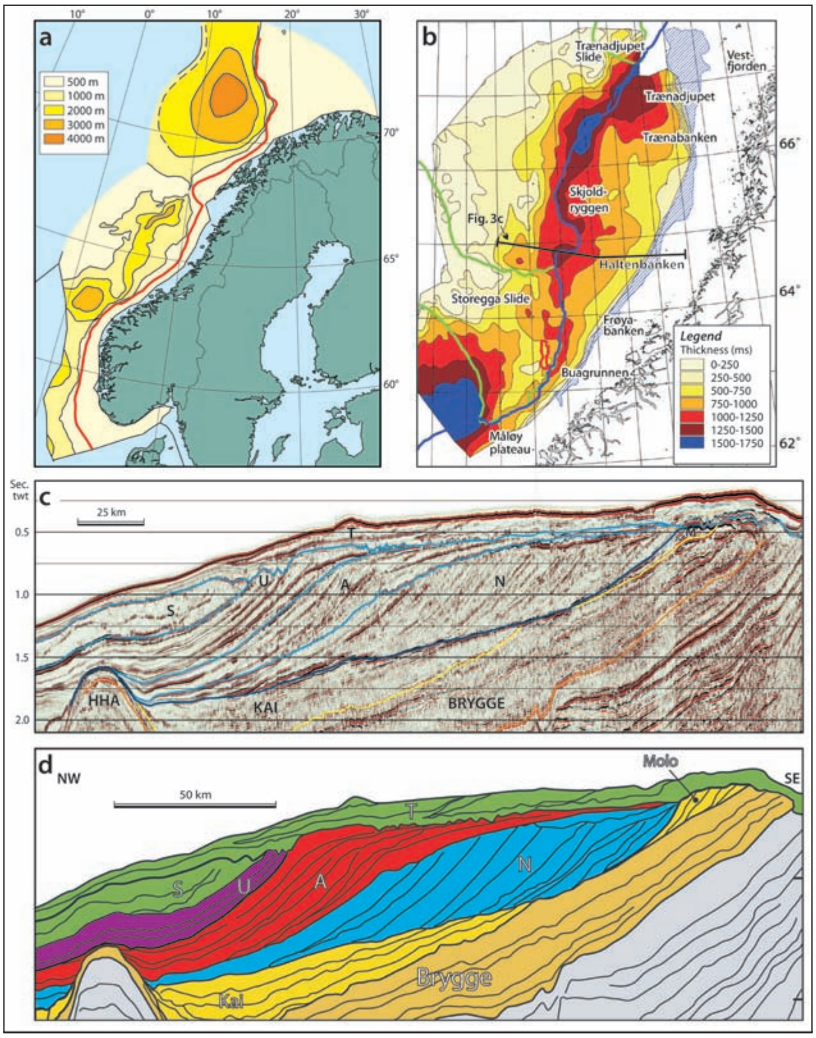


Figure 3. Distribution and thickness of the Quaternary (late Pliocene and Pleistocene) sediments along the Norwegian continental margin of Norway. (a) Seabed areas from the North Sea to the SW Barents Sea, modified from Riis (1996) and Vorren and Mangerud (2008), (b) the Mid-Norwegian shelf, modified from Rise et al. (2005), (c) seismic cross section from the Mid-Norwegian shelf, after Ottesen et al. (2007), and (d) interpretation of the profile in (c) with letters/names of formations/units. After Ottesen et al. (2007). Position of profile is indicated in (b).

to outline the development of the whole Mid-Norwegian shelf during the last 3 million years. For more details, we refer to the recent paper by Ottesen et al. (2009).

Land areas

The Quaternary stratigraphy of northern and central Norway has previously been compiled and presented as simplified logs on small-scale maps resulting from the so-called 'Nordkalott' and 'Mid-Norden' projects in northern and central Fennoscandia (Hirvas et al. 1988, Bargel et al. 1999). A modified version of selected parts of these data, with additional stratigraphical logs both from northern, central and southern Norway based on various sources, is presented here. The geological implications of these data are further considered in the accompanying paper by Olsen et al. (2013).

The distribution of various Holocene sedimentary deposits and formations that are mainly of nonglacial origin (deposits from rockfalls, landslides, fluvial deposits, weathering, shorelines etc.), is described by Fredin et al. (2013).

Overview of maps of Quaternary deposits onshore in Norway

Mapping of Quaternary sediments in Norway has been carried out during many decades, mostly in scales between 1:10,000 and 1:100,000 (see review by Bargel 2003). The surficial deposits are classified according to genesis. The legend and standard colour coding of the NGU Quaternary geological maps reflect various sediment types. However, it was not until the last three

Figure 4. Geographical position of most sites and areas mentioned in the text. For more details, see Appendix B, Figure B1 and Table B1, and for sites and areas not included here, see Figures 16, 17 and 33, and Appendix B (Figures B2–B5).

to four decades that mapping programs were initiated to produce detailed map sheets covering large areas. The maps were produced for many purposes, among others to locate building material for road construction, etc. The distribution of Quaternary sediments is important for many kinds of human activity, especially agricultural purposes and areal planning, but also for recreation (landscape development and geological history), education and research (ice-sheet modelling) purposes. Another important field of use of such maps is geological hazard or risk analysis for landslides, where, e.g., the distribution and thickness of marine clays are of vital importance.

Generalised maps of different regions have been produced at varying scales (Figure 5). These maps have been compiled into one generalised map covering all of Norway (http://www.ngu.no/kart/losmasse/), which has been simplified to represent merely sediment thickness and shown in reduced scale in Figure 1a. The original map scales were 1:500,000 for Finnmark county (Olsen et al. 1996b), 1:250,000 for Troms (Sveian et al. 2005), 1:400,000 for Nordland (Bargel 2001), and 1:250,000 for Sør-Trøndelag (Reite 1990), Møre & Romsdal (Follestad 1995), Sogn & Fjordane (Klakegg et al. 1989), Hordaland (Thoresen et al. 1995), Rogaland (Bergstrøm et al. 2010) and Aust-Agder (Riiber and Bergstrøm 1990). Furthermore, for Oslo and Akershus (Bargel 1997) and Østfold (preliminary version, Olsen et al. 2005) the generalised scales were 1:125,000, and for Jotunheimen 1:250,000 (Holmsen 1982) (for location, see Figures 4 and 5) and 1:1 million for the remaining parts of Norway (Thoresen 1990, Bargel et al. 1999). The different maps have been digitised, and the lines were imported into a Geographic Information System (GIS). Polygons were built from these lines, and each polygon was given a signature to the sediment type.

Our purpose in this paper is to give an overview of the Quaternary deposits by drawing and describing a simple, small-scale map (Figure 1a) based on a more complex Quaternary map (Figure 1b), where we clearly separate areas with exposed bedrock and bedrock with a discontinuous and thin (commonly <1 m) sediment cover from areas with a thick sediment cover. Therefore, the legend has been simplified to comprise only a few sediment types, all included in one category, which is dominated by till (Figure 1a). Exposed bedrock is pink and continuous cover of till (or sediments dominated by till) is green in both maps (Figure 1a, b). Certain areas have a relative continuous cover of organic deposits. However, these areas are not shown on the simplified map, but are instead given the colour of the underlying material (e.g., bedrock, till, etc.).

Sediment distribution and thickness in the different land areas

Generally, Norway has large areas of exposed bedrock or bedrock with a thin cover of Quaternary sediments. However, the southeastern parts of Norway (the lowland east and northeast of Oslo), areas adjacent to and under the former ice divide in southern Norway (Figure 6), the Jæren area in southwest Norway and Finnmarksvidda in northern Norway (Figures 1 and 4) have large areas with a generally continuous cover of sediments. The sediments are dominated by tills in most areas, but in the main valleys and basins waterlain sediments, including glaciofluvial, fluvial and marine deposits often represent the dominant sediment types. Based on natural sections, excavations, drillings and some seismic profiles, the average sediment thickness in areas with continuous cover of sediments (e.g., at Finnmarksvidda) is estimated to *c.* 6 m, similar to northern and central Finland and Sweden (The Nordkalott Project 1986, The Mid-Norden Project 1999). However, in some valleys and basin areas dominated by marine deposits the sediment thickness is much thicker, commonly more than 10 m, and occasionally the sediment thickness is more than one hundred metres, e.g., up to 400 m sediment thickness in the northernmost, outer part of Gauldalen near Trondheim.

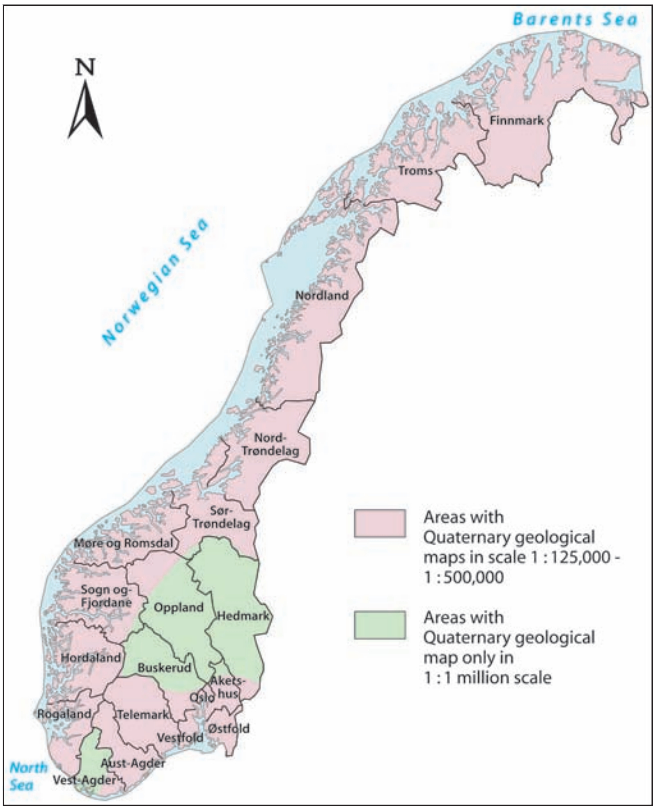


Figure 5. Counties in Norway, most of them mentioned in the main text. Counties with existing Quaternary geological maps in various scales are indicated.

The lowland east and northeast of Oslo (Østlandet)

The fields with continuous sediment cover east and northeast of Oslo are all dominated by marine deposits within the nearest 50 km. These areas are therefore situated mainly below the postglacial marine limit (*c.* 200 m a.s.l.) and have a long history of Holocene fluvial erosion and clay slides. This has changed the topography of the landscape rapidly and dramatically within minutes or hours at certain times in the landscape evolution history. The largest deposits of sand and gravel in Norway, which is mainly of glaciofluvial origin, are located in the northeastern lowland (e.g., Gardermoen airport and surrounding areas).

The sediment thickness in this area may locally reach more than 100 m, e.g., in the Hauerseter area where seismic measurements indicate up to 120–125 m sediment thickness (Østmo 1976, Longva 1987).

The central inland of SE Norway—the ice divide and surrounding areas

During initial phases of glaciations ice growth expanded from core areas along the regional water divide in the Scandinavian mountains, and during phases of maximum glaciations the ice divide migrated to a more easterly and southeasterly position (Figure 6). During the last glaciation, the most extreme easterly and southeasterly position of the ice divide was at, or south of Vinstra in Gudbransdalen (Figures 4 and 6), but all recorded field data suggest that the E–W-trending ice divide maintained a position north of Lillehammer at the southern end of Gudbrandsdalen also during the Last Glaciation Maximum (LGM) (Olsen 1983, 1985a, b). It is clear from the map (Figure 1), that large areas in central southeast Norway (along the former ice divides) are characterised by a continuous sediment cover. This zone extends some 50–80 km beyond the LGM ice-divide zone both in the north and south, which indicates that the glacial erosion, which is generally strong closer to the glacier margins, has been significantly reduced already at a relatively long distance from the ice divide zone.

The sediment thickness in these central inland areas is often more than 5 m on the plains and more than 10 m in the small valleys. In larger valleys, such as Gudbrandsdalen and lower parts of Østre Gausdal the sediment thickness (mainly till) may reach more than 100 m (e.g., Olsen 1985a).

The Jæren area

The Jæren area (Figure 4) (Appendix B, Figure B1: 107, *c.* 25x50 km^2) in southwestern Norway has generally a continuous sediment cover. In the southern coastal part a sediment thickness of more than 100 m is recorded, e.g., at Grødeland (Janocko et al. 1998). The main part of these sediments is till, with some intercalated marine or glaciomarine sediments. At higher elevations (up to more than 200 m a.s.l.) and some 6–10 km from the coast in the southern parts of Jæren (Høgjæren), several-metre thick marine sediments of Middle Weichselian age are recorded at many sites (e.g., Andersen et al. 1987, Larsen et al. 2000). This is considerably higher than the late-/postglacial marine limit, which rises from 0 to 30 m (a.s.l.) northwards in the Jæren area (Andersen et al. 1987).

Finnmarksvidda

Finnmark County in northern Norway has together with Hedmark County in the southeast (Figure 5), the largest fields with continuous sediment cover in Norway. The main part of Finnmarksvidda is located south of the fjord heads. It is a lowland plateau, *c.* 300–500 m a.s.l. and more than 10,000 km^2, with some mountain peaks reaching at least 1000 m a.s.l. in the north. The sediment thickness, which is dominated by tills in this area is estimated to *c.* 6 m in average (Olsen 1988), but may in some cases reach more than 50 m (Olsen et al. 1996a). In the delta areas along the major rivers, e.g., at the Tana River delta the sediment thickness is up to at least 100 m (Mauring and Rønning 1993, Mauring et al. 1995).

Other parts of Norway

Continuous and/or thick Quaternary deposits in areas of size at least tens of square kilometres occur also at Lierne in Nord-Trøndelag. In all other parts not mentioned separately the Quaternary sediments are scarce or only present as thin or discontinuous 'blankets' on bedrock.

Sediment distribution and thickness on the shelf

The North Sea Plateau

The North Sea Plateau is a shallow shelf area south and west of the Norwegian Channel (Figures 4 and 6) that has generally been outside or at the margins of the Fennoscandian/ Scandinavian ice sheet. The ice-sheet configuration in these shelf areas is generally poorly known, with large uncertainties during the most extensive glaciations that occurred during the last 0.5 million years and also during older glaciations in Scandinavia. Studies of seismic profiles show a complicated history with extensive erosion over large areas. This is especially in the eastern areas, in a zone where the westward-expanding Scandinavian ice sheet has reached during several glaciations. In these areas sediment-filled, deep erosional channels are abundant in the Pleistocene sediments, and the channels have been interpreted as subglacial formations (e.g., Sejrup et al. 1991). In the central parts of the North Sea Basin thick sequences (adding to *c.* 1 km thickness) of Late Pliocene–Pleistocene sediments, thickening from the east towards the central area in the west exist (Eidvin et al. 1999). These are mainly deposited through glaciofluvial and glaciomarine processes during the last three million years. However, the age control for these sediments is poor, implying a big uncertainty in the volume estimates.

The Mid-Norwegian shelf

Two major depocentres exist on the Mid-Norwegian shelf. In the north (65°–67°N), large areas along the present shelf edge have more than 1000 m of glacial sediments, whereas at the mouth of the Norwegian Channel outside Stad (62°N) a large trough-mouth fan (TMF) has been deposited. The sediment thickness of the fan may exceed 1500 m. The deltaic Molo Formation that has recently been redated to a Mid/Late Miocene to Early Pliocene age (Eidvin et al. 2007), is located parallel to the Norwegian coastline (Figure 3b, c), generally 60–80 km west of it. The prograding Pliocene/Pleistocene wedges start outside the Molo Formation (Figure 3c, d). Seismic profiles show that the Mid-Norwegian shelf has prograded up to 150 km towards the west, and more than 1000 m of glacially derived sediments have been deposited during the

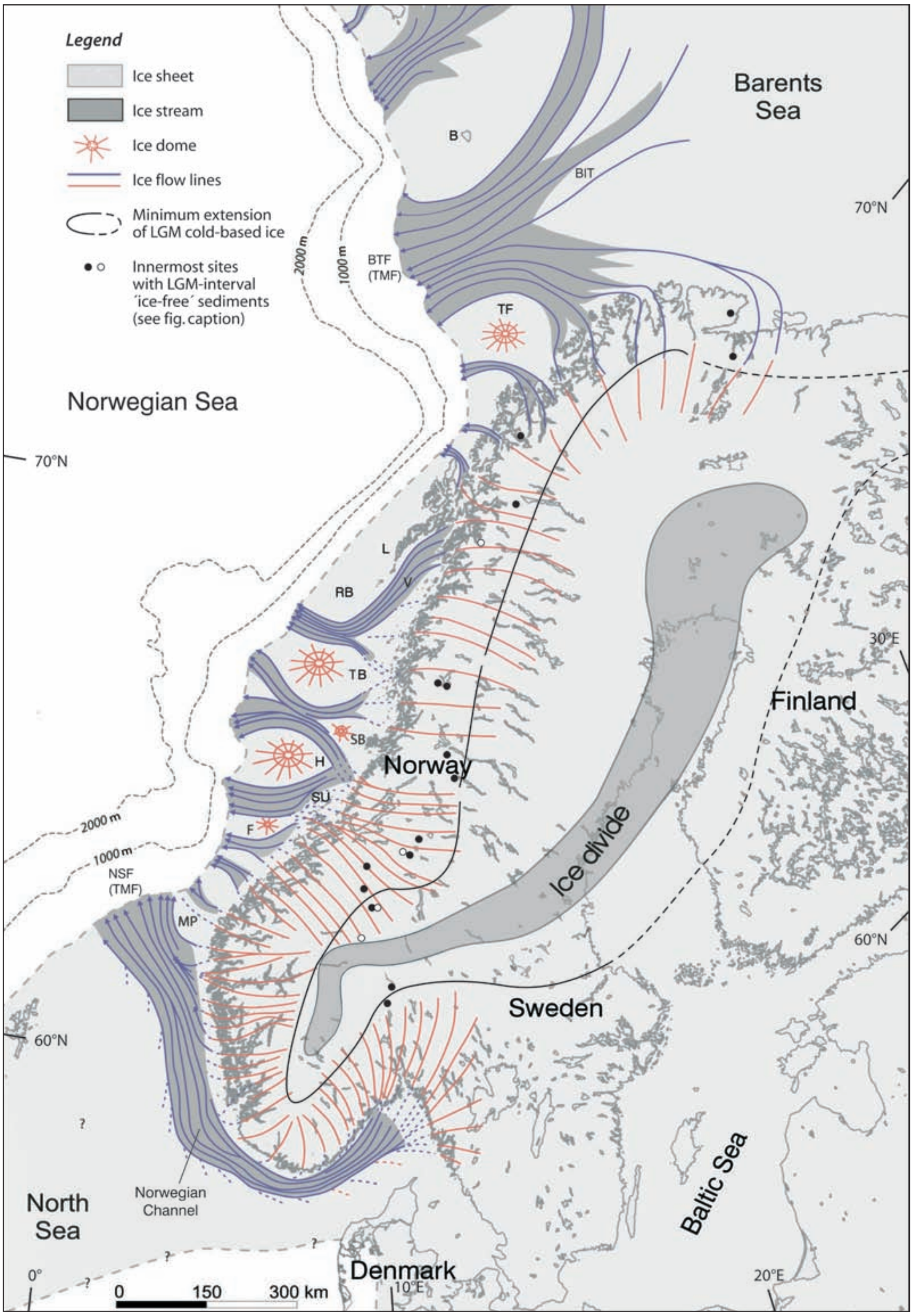


Figure 6. Large-scale ice-flow model of the western part of Scandinavia during the last glacial maximum, modified from Ottesen et al. (2005a). Minimum areas (partly discontinuous and stippled line). with ice frozen to the underground (after Kleman et al. 1997) and innermost locations of sites with sediments deposited during ice-free conditions or proglacially during ice-marginal retreat in the LGM interval are marked (after Olsen et al. 2002). Type of material and dating method are indicated (dot indicates ^{14}C-dated bulk plant remains, and open ring indicates ^{14}C-dated animal bones, U/Th-dated calcareous concretions or OSL-dated sand). B= Bjørnøya (Bear Island), BIT= Bear Island Trough, BTF= Bear Island Trough Fan, TMF= Trough-Mouth Fan, TF= Tromsøflaket, V= Vesterålen, L= Lofoten, RB= Røstbanken, VF= Vestfjorden, TB= Trænabanken, H= Haltenbanken, SU= Sularevet, F= Frøyabanken, MP= Måløyplatået, NSF= North Sea Fan.

last 3 million years on the shelf. For more details about the Mid-Norwegian shelf, see, e.g., Ottesen et al. (2009).

The Vestfjorden/Lofoten/Vesterålen areas

The Lofoten/Vesterålen shelf (67º–69ºN) is generally narrow (Figures 2 and 4) with a thin cover of Quaternary sediments (generally less than 20 m). The largest sediment thickness (>100 m) is found along the shelf edge and in the upper slope. Due to the steep slope with marked valleys and slide scars, a good estimate of the total volume of Pliocene/Pleistocene sediments is difficult to elaborate. The sediment volume may be difficult to estimate also in the Vestfjorden–Trænadjupet area (Figure 7), but many megascale lineations and large grounding-line moraines indicate fast-flowing glaciers as an effective erosion and transport agent. The lineations are several kilometres long ridges, up to 10 m high and 200–500 m wide, and are separated by 300–700 m-wide trenches, whereas the grounding-line moraines are up to several tens of metres thick (Ottesen et al. 2005b).

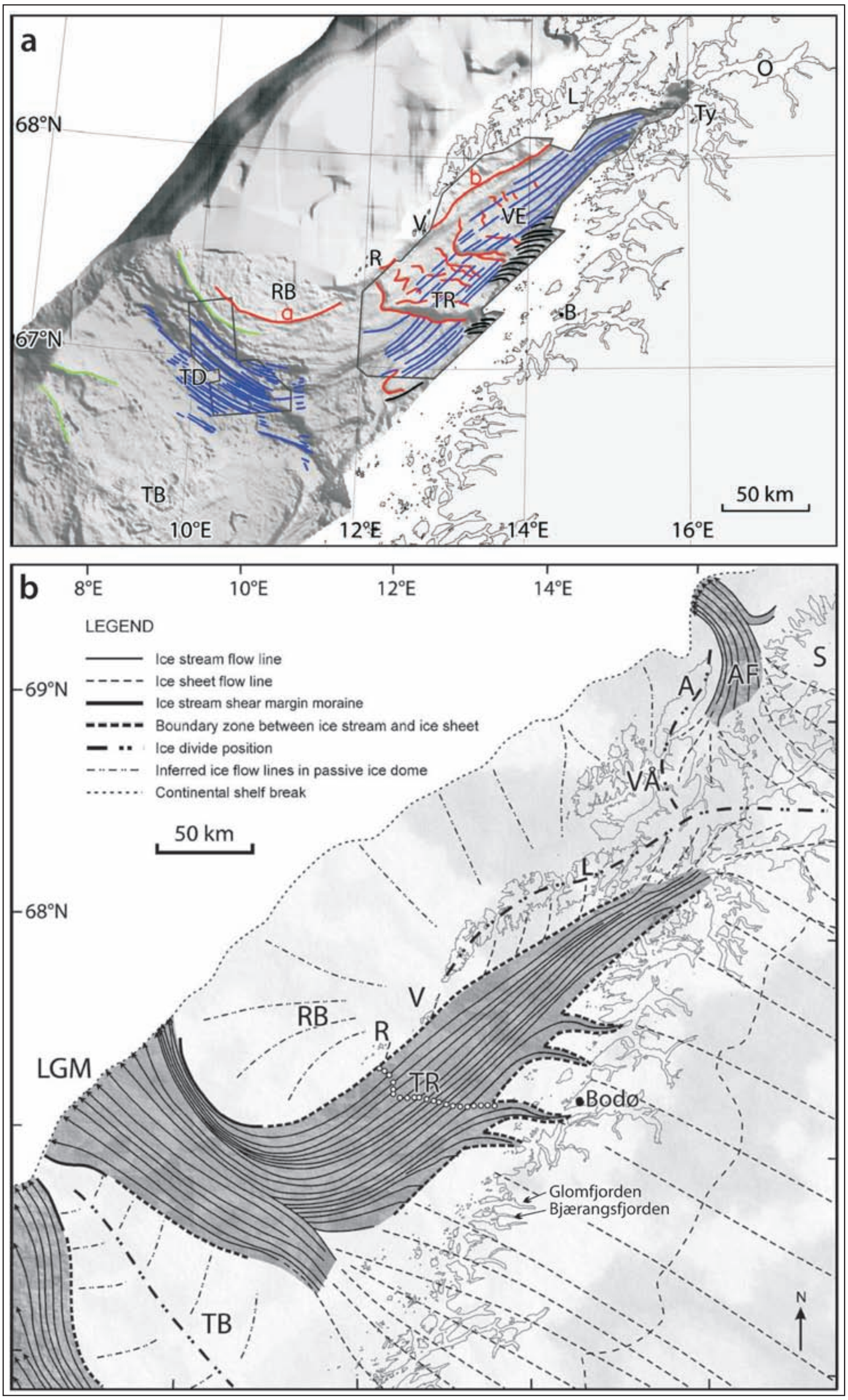


Figure 7. (a) Sea-floor morphology of the Vestfjorden/Trænadjupet with ice-stream marginal and grounding-line moraines (red lines) and megascale glacial lineations (MSGL, blue lines) indicating fast ice flow, and (b) ice-flow model of the Lofoten/Vesterålen area and adjacent shelf. Modified from Ottesen et al. (2005b). Legend of letters: O= Ofotfjorden, Ty= Tysfjorden, L= Lofoten, VE= Vestfjorden, V= Værøy, R= Røst, TR= Tennholmen Ridge, B= Bodø, RB= Røstbanken, TD= Trænadjupet, and TB= Trænabanken. Additional legend for b: S= Senja, AF= Andfjorden, A= Andøya, and VÅ= Vesterålen.

The northern shelf areas off Troms and in the southwest Barents Sea

The Troms shelf is narrow (35 km wide) in the south (69°30'N) outside Senja (Figure 4), but widens towards the north into the Barents Sea. The shelf comprises shallow bank areas with intermediate cross-shelf troughs. The main transport of glacial debris probably followed these depressions, which have hosted the palaeo ice streams whereas the shallow bank areas were covered by passive ice (domes) (Figure 6). Due to the steep slope (up to 10°) outside the shelf break off Troms, it is difficult to estimate the volumes of the glacial debris being transported into the deep sea. Generally, there is a sparse cover of Quaternary sediments on the banks, highly variable but mostly less than 25 m, and certain areas may completely lack Quaternary sediments (Rokoengen et al. 1979).

Tromsøflaket (20,000 km^2) is a large, shallow bank area (water depth 150–350 m) on the northern Troms shelf at the transition to the Barents Sea. This large area was probably in some intervals covered by a passive ice dome, which directed ice flow south and north of the area. Most of the ice from Finnmark was drained out the fjords, then being deflected towards the west, and farther northwards into the Bear Island Trough were it coalesced with the main trunk of the Bear Island Trough ice stream. The Bear Island Fan is a huge trough-mouth fan at the shelf edge outside the Bear Island Trough (125,000 km^2) with a sediment thickness of up to 4 km (Figure 3a, Riis 1996, Vorren and Laberg 1997).

During the Cenozoic, strong glacial erosion of the floor of the Barents Sea has developed a very marked angular unconformity over large areas. Up to 300 m of glaciogenic sediments overlie this unconformity (Sættem et al. 1992, Vorren et al. 1998, Lebesbye 2000). These sediments can be separated into seismic units separated by smooth or irregular surfaces. Each of the sequences can be up to 150 m in thickness (Lebesbye 2000). Lebesbye mapped a complex sequence of glaciogenic sediments with up to 250 m in thickness in the areas outside the Finnmark coast, revealing alternating phases of deposition and erosion. He also identified several glacial phases and related most of the seismic units to Weichselian or Saalian age. The depocentre developed in a buried wide glacial trough running parallel to the coastline (Lebesbye 2000).

Zones of major erosion and accumulation during the Quaternary

The surface-near sediments of Norway and adjacent Nordic countries comprise mainly sediments from the last glaciation, the Weichselian. However, these young sediments in large areas in the inland of southeast Norway, the Jæren area in southwest Norway, and parts of Finnmark in northern Norway are commonly underlain by deposits from preceding glacials or interglacial/interstadial periods. This is also the case for even larger areas in Sweden and Finland (Hirvas et al. 1988, Bargel et al. 2006).

It has been compiled a till stratigraphy of Finnmark, outlining up to five superimposed till beds with interlayered interglacial/interstadial sediments (Olsen 1988, 1993a, b, Olsen et al. 1996a). Similar compilations are carried out for northern and central Finland and Sweden (The Nordkalott Project 1986, The Mid-Norden Project 1999).

On the basis of the stratigraphical studies in Finnmark it has been mapped certain areas with weathered bedrock in situ, indicating only minor or moderate glacial erosion (Figure 8, Olsen 1998). In Finnish Lapland large areas with preglacial weathered bedrock have been recorded (Hirvas 1991). The same is also registered in northern Sweden (The Nordkalott Project 1986) and in several parts of Norway where also blockfields occur frequently (Figure 8, Thoresen 1990).

The areas under the central parts of the Fennoscandian palaeo ice sheet have generally a continuous sediment cover (Hirvas et al. 1988, Thoresen 1990, Bargel et al. 1999). In these areas ice probably was frozen to the ground during a substantial part of each major glacial cycle, in contrast to the areas closer to the ice margin (e.g., Kleman et al. 1997, Boulton et al. 2001). This is indicated for the last glaciation by the occurrence of numerous multiphase deglaciation features, e.g., crossing patterns of lateral and sublateral drainage channels, all represented in the same area and with a diversity that could not derive from only one deglaciation phase (Figure 9) (Kleman 1990, Sollid and Sørbel 1994, Follestad 2005c, Follestad and Fredin 2007).

The western part of the Fennoscandian ice sheet was efficiently drained through a number of palaeo ice streams (Figure 6). The erosion of the Norwegian land areas was extensive, and during the glacial periods most fjords were eroded and emptied for all sediments (e.g., Aarseth et al. 1997), that generally had been deposited during the retreat of the previous glacial and the subsequent ice-free period. A large amount of the glacially derived sediments were transported onto the shelf and beyond the shelf edge at peak glaciations.

Recently, Kleman et al. (2007) have described and classified patterns of Quaternary ice-sheet erosion and deposition in Fennoscandia, and they have suggested a simple, logical, and quite useful theoretical framework for explanation. The Scandinavian mountain-centred ice sheets and the full-sized Fennoscandian ice sheets, which here and in the accompanying paper

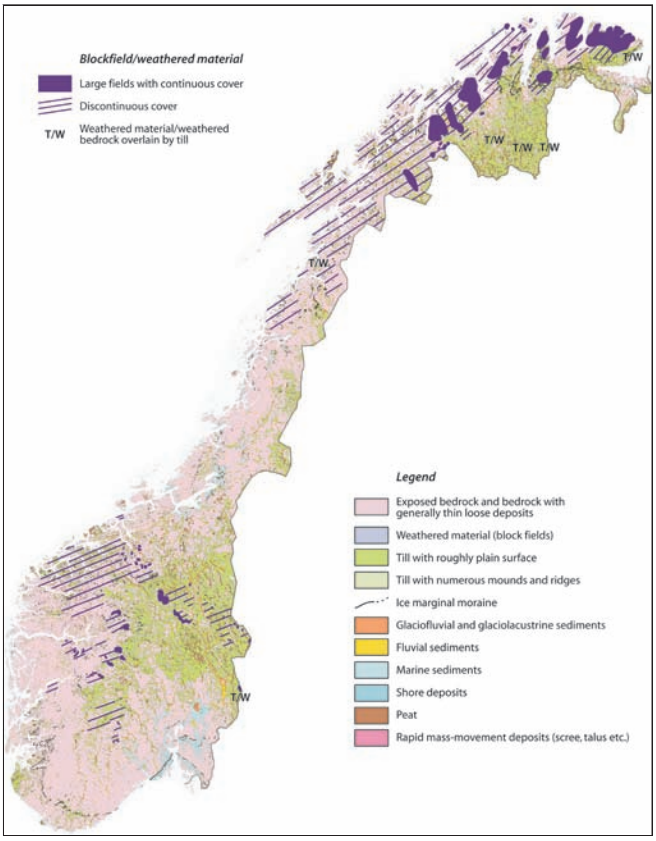


Figure 8. Sites and areas characterised by weathered material (mainly blockfields) and/or weathered bedrock, with or without a Quaternary sedimentary cover, modified from Thoresen (1990).

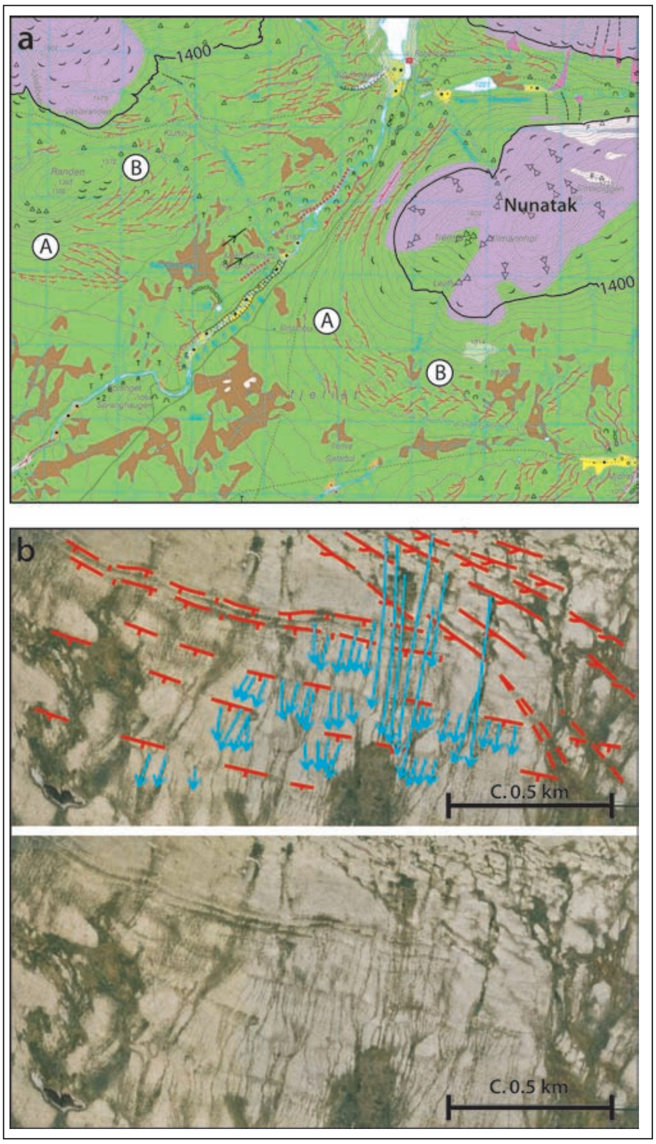


Figure 9. (a) Part of Quaternary geological map Rondane 1718 I (scale 1:50,000) with two sets (A, drainage towards east-southeast, B, drainage towards southwest) of crossing lateral drainage channels, from two different deglaciation phases separated by at least one phase with a cold-based ice cover, after Follestad (2006). Areas higher than 1400 m a.s.l. (indicated) may have been nunataks during the last deglaciation after mid Younger Dryas. North is up and each coordinate net square (light blue lines) on the map represent 1x1 km^2 in the terrain. (b) Air-photo of an area adjacent to the letter A, to the left in a. Example of several sets of lateral meltwater channels, from central southeast Norway. The drainage shifted between a NE-trending to a SW-trending direction several times during the represented deglaciation history. Several downslope directed meltwater channels (blue colour), many of these thought to derive from subglacial groundwater flow (cf., Gjessing 1960), are represented and are grouped in sets starting in lateral positions at different elevations. These features indicate multiple phases with transitions from cold-based to warm-based ice. Photograph copied from internet (www.norgeibilder.no).

by Olsen et al. (this issue) are described as ice sheets of S (or SIS) and F (or FIS) configurations, respectively, are denoted as MIS[2]-style and FIS-style ice sheets by Kleman et al. (2007). They concluded, for example that the western (fjord) zone of deep glacial erosion (Figure 10) formed underneath both SIS- and FIS-style ice sheets during the entire Quaternary, whereas the eastern (lake) zone of deep glacial erosion is exclusively a product of SIS style ice sheets, and formed mainly during the early and middle Quaternary. They also found that the eastern zones of scouring misfit with respect to SIS configuration, whereas the thick (mainly central and northern Sweden) drift zone (Figure 10) is interpreted to be a result of more than 1.1 million years of SIS-style ice sheet eastern margin locations and drift deposition on the Scandinavian peninsula, and of a subsequent survival of these drift deposits underneath central low-velocity and cold-based parts of FIS-style ice sheets during the last 1 million years.

Quaternary stratigraphical data from Norway and adjacent land/seabed areas

In contrast to the abundance of information which is available from the last glaciation (Weichselian) in Fennoscandia, terrestrial data from previous glaciations in this area are sparse and rely on very few sources of information. The interpretation and correlation of data from pre-Weichselian glaciations follow the same principles as those from the last glaciation and rely to a large extent on the validity of these. Tills from the Elsterian (MIS 12, i.e., marine isotope stage 12) (see e.g., Olsen et al., 2013, fig. 5) (Šibrava 1986, Lowe and Walker 1997) or previous glaciations have not so far been reported or confirmed from onshore positions in Fennoscandia, although data from the North Sea, the Norwegian Sea, the Netherlands, Germany and Eastern Europe suggest very strongly that the ice cover must have included, and in fact been initiated from the mountainous regions of Fennoscandia/ Scandinavia several times during the pre-Saalian (sensu stricto). In this compilation, only the glaciations directly traced in Fennoscandia are included, except to some old glacial deposits occurring on the adjoining shelf in the west, which are also mentioned briefly.

It is necessary to comment on the widely used term Saalian as the name for the penultimate glaciation in Northern Europe. Traditionally this term was used for the whole period between the Holsteinian and Eemian interglacials, but only the last part of the period, probably correlating with the marine isotope stage (MIS) 6, was fully glaciated (the Drenthe and Warthe stades). The long initial part of the Saalian (sensu lato), which alternated between cold and temperate intervals in Germany and the Netherlands, is now known to include glacial intervals, as well as interglacials and interstadials in northern Europe as a whole (Appendix A, Table A1, see also Appendix B, Tables B2 and B3, and Olsen et al., 2013, fig. 2).

Both terms Saalian (sensu stricto) and Late Saalian (sensu

[2] We prefer not to use the abbreviation MIS with respect to ice-sheet configuration since it is presently more often used as an abbreviation for marine isotope stage (Šibrava 1986, Lowe and Walker 1997) which is referred to several times in this and the accompanying papers (this issue).

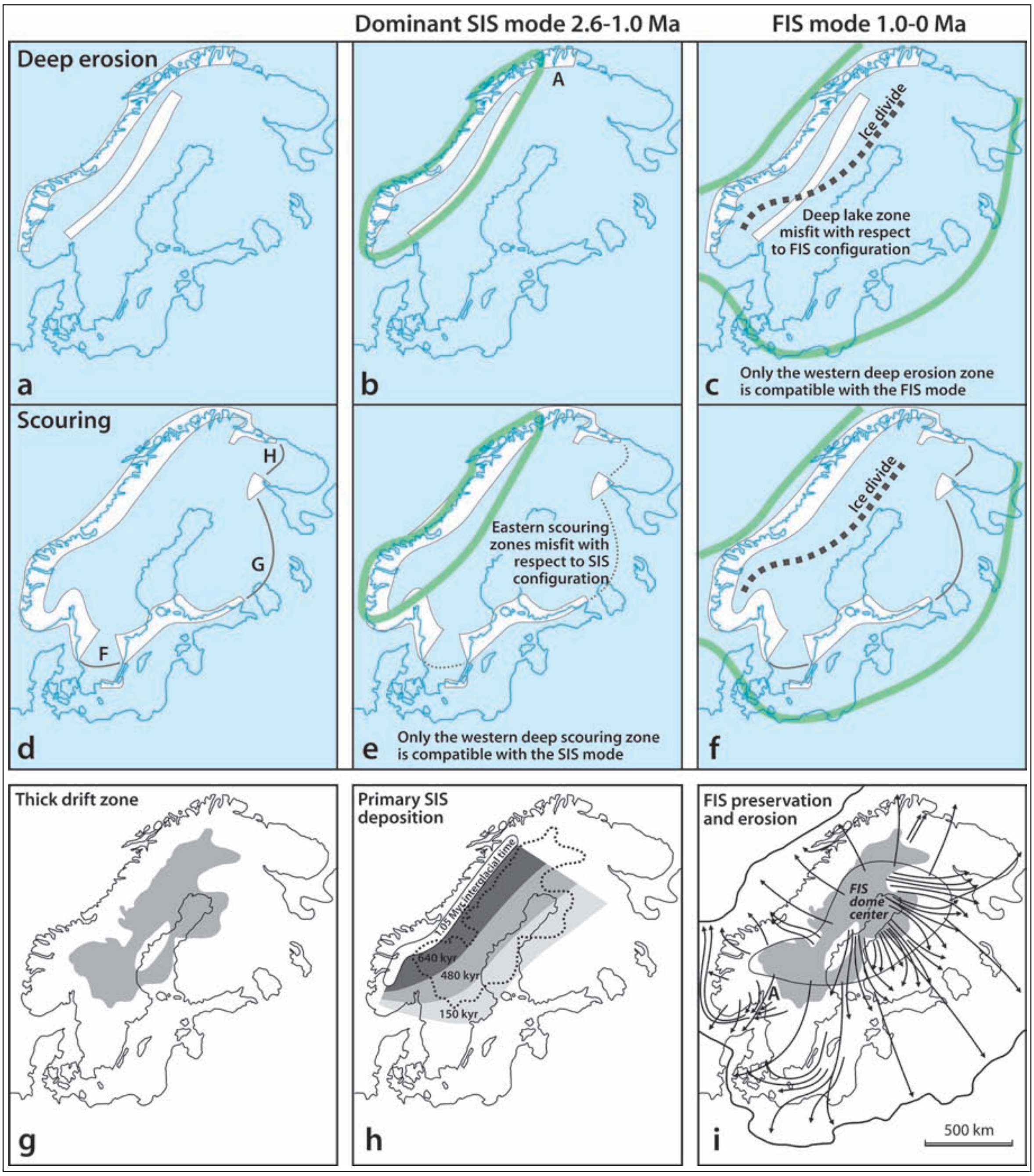


Figure 10. Combined Figures 10 and 11 from Kleman et al. (2007), slightly modified. The maps show patterns of Quaternary ice-sheet erosion and deposition in Fennoscandia. Published with permission from Elsevier and Copyright Clearance Center's Rightslink service, license number: 2075350071489, Nov. 24, 2008.

lato) are here maintained as the name for the penultimate glaciation, of MIS 6 age, whereas the previous long complex period, which followed the Holsteinian interglacial is subdivided in a Saalian Complex, including Early and Middle Saalian cold intervals (stadials) separated by a warmer interval (interglacial or interstadial) of MIS 9 age, and a proper interglacial correlating with MIS 7.

The represented ages and geographical distribution of all Quaternary stratigraphical sites used here, showing that all main regions are represented, are listed in Table B4 (Appendix B).

Principles for stratigraphical correlation used in northern and central Fennoscandia

The principles of correlation in central and northern Norway follow those presented and used in the Nordkalott- and the Mid-Norden projects, which were based on collaboration between the Geological Surveys of the Nordic countries during the 1980s and 1990s (e.g., Hirvas et al. 1988, Bargel et al. 1999). The clast fabric, provenance and lithology, and thereby the associated ice-flow direction associated with each till bed, are used as the main tools for correlation from site to site (Olsen 1988, 1993b, Olsen et al. 1996a). The age assignments are

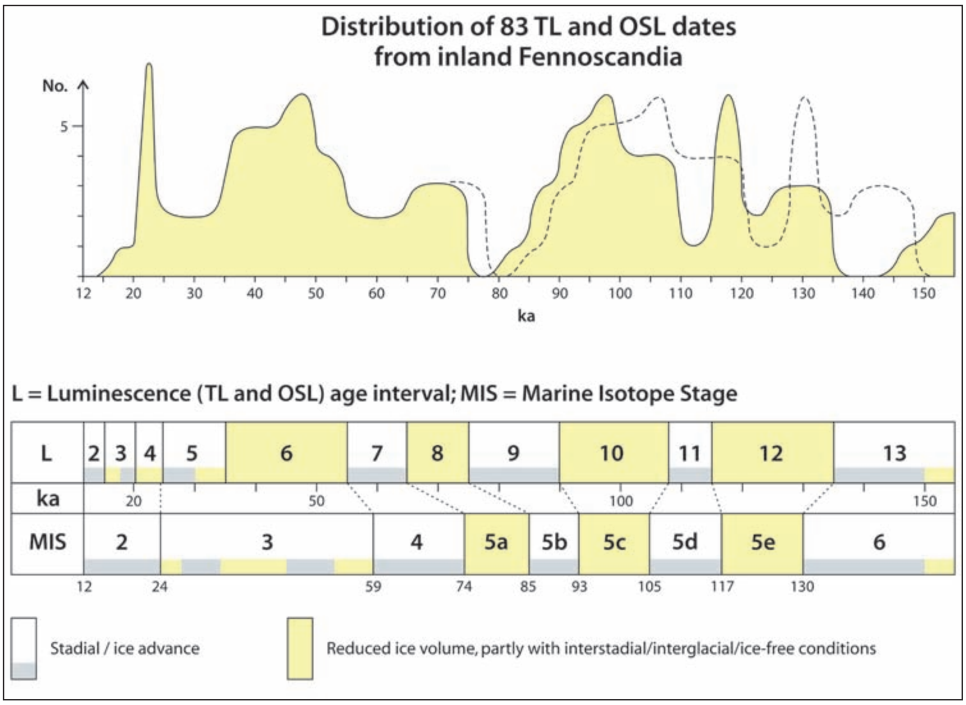


Figure 11. Frequency distribution of 83 luminescence dates of sediments from the inland of Fennoscandia. Most of these dates are from Olsen et al. (1996a), but are here used without correction for shallow traps since the argument for such traps was based on data from Greenland and Denmark and is not (shown to be) valid for Fennoscandia (Olsen, in Lokrantz and Sohlenius 2006). An alternative curve is indicated (stippled) for the possibility of a general underestimation of c. 10% of older ages, >75–100 ka. Seven recently reported OSL dates from subtill glaciofluvial/fluvial sediments from the inland of Trøndelag are also included (after Johnsen et al. 2012).

mainly based on dates from the individual sites, correlations to other dated stratigraphies, and comparison with a general age model (Figure 11) defined by the distribution of luminescence dates (TL and OSL) of sediments from intercalated interstadials/ice retreats from the central inland areas of Fennoscandia. For further details on correlations and correlation principles we refer to Appendix A.

Quaternary stratigraphies and correlations during major (F style) glaciations

Deposits from major pre-Eemian substages, interstades and interglacials

A record of 11 onshore Norwegian sites (Figure 12) with pre-Eemian deposits of glacial, interglacial and interstadial origin is compiled in Table B5 (Appendix B). These and many similar sites from adjacent onshore and offshore areas are briefly described in the following text.

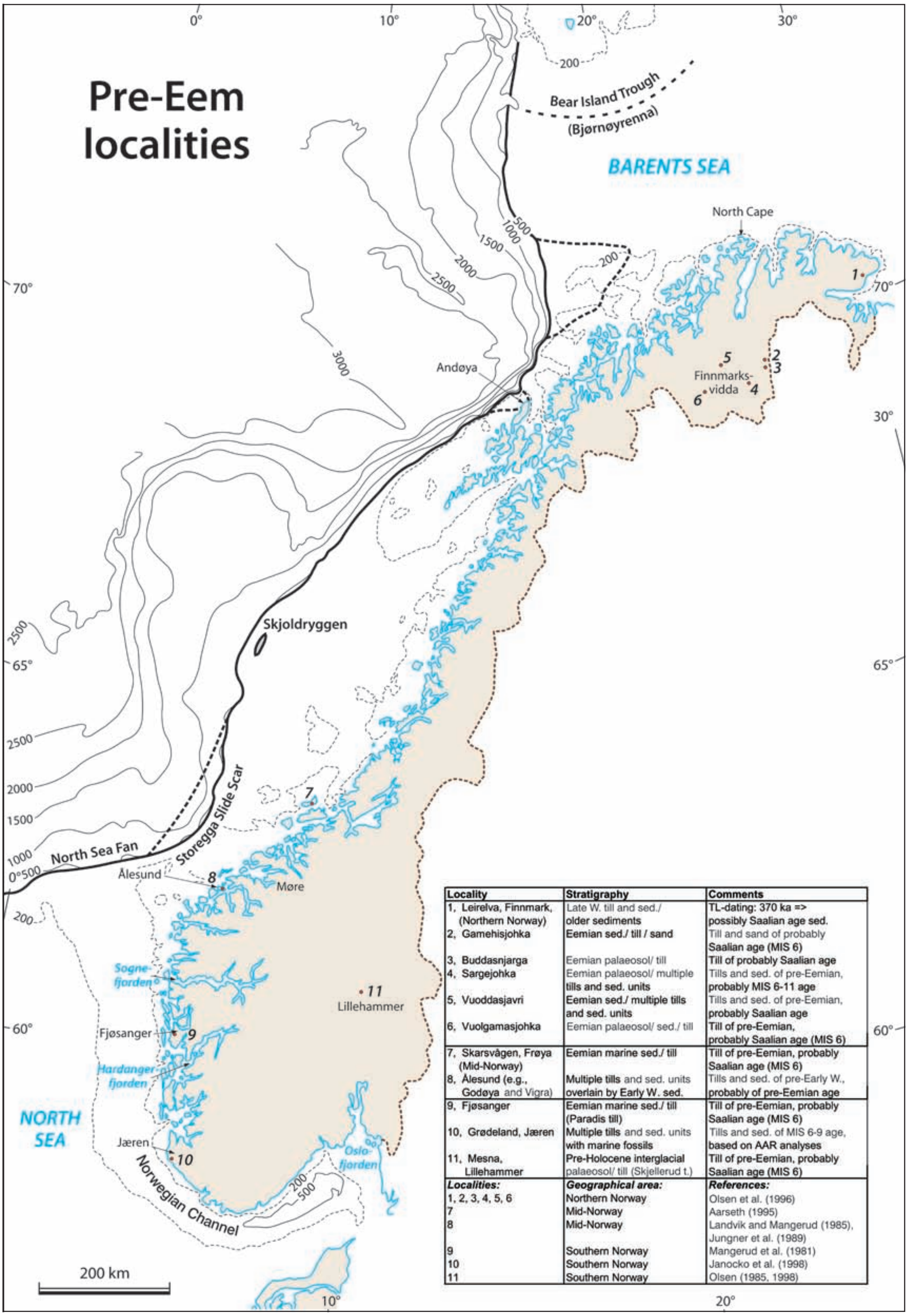
Northern to central Norway and adjacent shelf and land areas In northern Fennoscandia a record of 17 Finnish, 18 Swedish and 5 Norwegian localities with pre-Weichselian till beds have been reported (The Nordkalott Project 1986: Map of Quaternary geology, sheet 4, Olsen 1993b). Farther south, in central Fennoscandia, 21 Finnish localities, no Swedish and several offshore Norwegian localities are recorded (The Mid-Norden Project 1999: Map of Quaternary stratigraphy, Dahlgren 2002). Examples with stratigraphical information from some of these localities are briefly described in the following text.

Northern and central Norway. In Finnmark, northern Norway two localities with pre-Saalian (sensu stricto) tills and five with Saalian tills have been described (Figure 13 and Appendix B, Figure B2) (Olsen et al. 1996a). Most of these localities are located on eastern Finnmarksvidda, and according to (one of) the used criterions (Appendix A) this implies associated glacier extensions reaching at least to the coastal areas of northern Fennoscandia.

The relatively long and complex stratigraphy of tills and intercalated waterlain glacial and/or nonglacial sediments at Vuoddasjavri and Sargejohka, located at central and eastern Finnmarksvidda, respectively, comprise two of the most important datasets for reconstructing the pre-Weichselian glacier fluctuations in the northern Fennoscandia (Olsen 1993b, 1998, Olsen et al. 1996a).

The oldest sediment unit, Sargejohka unit K (Figure 14) is inferred to be of glaciofluvial origin, and simply based on the location of Sargejohka it derives most likely from the deglaciation of an ice sheet that had a maximum extension reaching at least to a position similar to the early Preboreal–late Younger Dryas glacier margins in the inner fjord areas in the north. Based on counting from the top, stratigraphical relationship to other recorded regional events (e.g., Appendix B, Table B2) and luminescence dates from younger units, the palaeosol (p6) that is developed in unit K is suggested to correlate with the Holstein interglacial (MIS 11), although faunal and vegetational evidence of warm conditions are so far not reported (Olsen 1998). This implies that unit K may represent the retreat of an ice sheet of MIS 12 age (or older). The overlying tills representing Glaciations 7–4 include ice-directional indications that imply S or F ice-sheet (SIS or FIS) configurations (Figure 15, see also Olsen

Figure 12. Pre-Eemian sites (MIS 6 – MIS 11, i.e., about 140–400 ka).



Locality	Stratigraphy	Comments
1, Leirelva, Finnmark, (Northern Norway)	Late W. till and sed./ older sediments	TL-dating: 370 ka => possibly Saalian age sed.
2, Gamehisjohka	Eemian sed./ till / sand	Till and sand of probably Saalian age (MIS 6)
3, Buddasnjarga	Eemian palaeosol/ till	Till of probably Saalian age
4, Sargejohka	Eemian palaeosol/ multiple tills and sed. units	Tills and sed. of pre-Eemian, probably MIS 6-11 age
5, Vuoddasjavri	Eemian sed./ multiple tills and sed. units	Tills and sed. of pre-Eemian, probably Saalian age
6, Vuolgamasjohka	Eemian palaeosol/ sed./ till	Till of pre-Eemian, probably Saalian age (MIS 6)
7, Skarsvågen, Frøya (Mid-Norway)	Eemian marine sed./ till	Till of pre-Eemian, probably Saalian age (MIS 6)
8, Ålesund (e.g., Godøya and Vigra)	Multiple tills and sed. units overlain by Early W. sed.	Tills and sed. of pre-Early W., probably of pre-Eemian age
9, Fjøsanger	Eemian marine sed./ till (Paradis till)	Till of pre-Eemian, probably Saalian age (MIS 6)
10, Grødeland, Jæren	Multiple tills and sed. units with marine fossils	Tills and sed. of MIS 6-9 age, based on AAR analyses
11, Mesna, Lillehammer	Pre-Holocene interglacial palaeosol/ till (Skjellerud t.)	Till of pre-Eemian, probably Saalian age (MIS 6)
Localities:	***Geographical area:***	***References:***
1, 2, 3, 4, 5, 6	Northern Norway	Olsen et al. (1996)
7	Mid-Norway	Aarseth (1995)
8	Mid-Norway	Landvik and Mangerud (1985), Jungner et al. (1989)
9	Southern Norway	Mangerud et al. (1981)
10	Southern Norway	Janocko et al. (1998)
11	Southern Norway	Olsen (1985, 1998)

et al., this issue) with associated ice extensions reaching to the coast, the shelf or even to the shelf edge.

Stratigraphical information from the pre-Weichselian glacial deposits from Vuoddasjavri and Sargejohka, which is correlated with other sites from Finnmarksvidda (Figures 12 and 13) and synthesised with regard to glacier extension is presented in Appendix B, Table B3 (see Substages 9–13, with comments in the north–column of maximum ice extension).

In central Norway very few Quaternary deposits of pre-Weichselian age are reported. Resedimented spruce pollen and

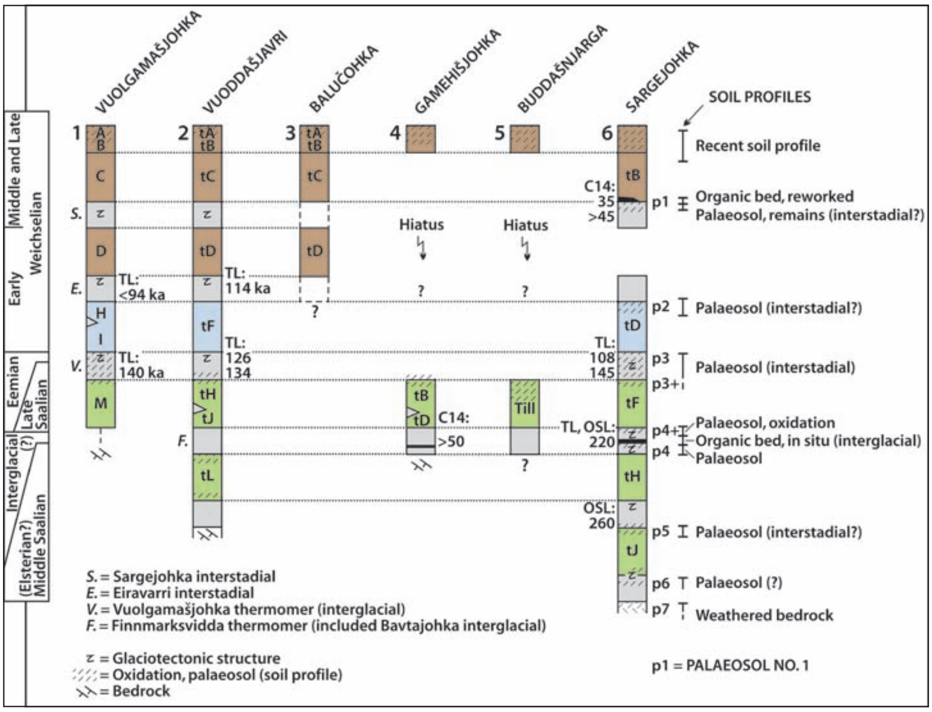


Figure 13. Six correlated Quaternary stratigraphies from Finnmarksvidda, modified from Olsen et al. (1996a). Localities 1–6 are shown in Figure B1 (Appendix B), as localities 15, 16, 12, 10, 11 and 13, respectively. The location of five of these are also included in Figure 12 (with other locality numbers, 1= 6, 2= 5, 3= not represented, 4= 2, 5= 3, and 6= 4).

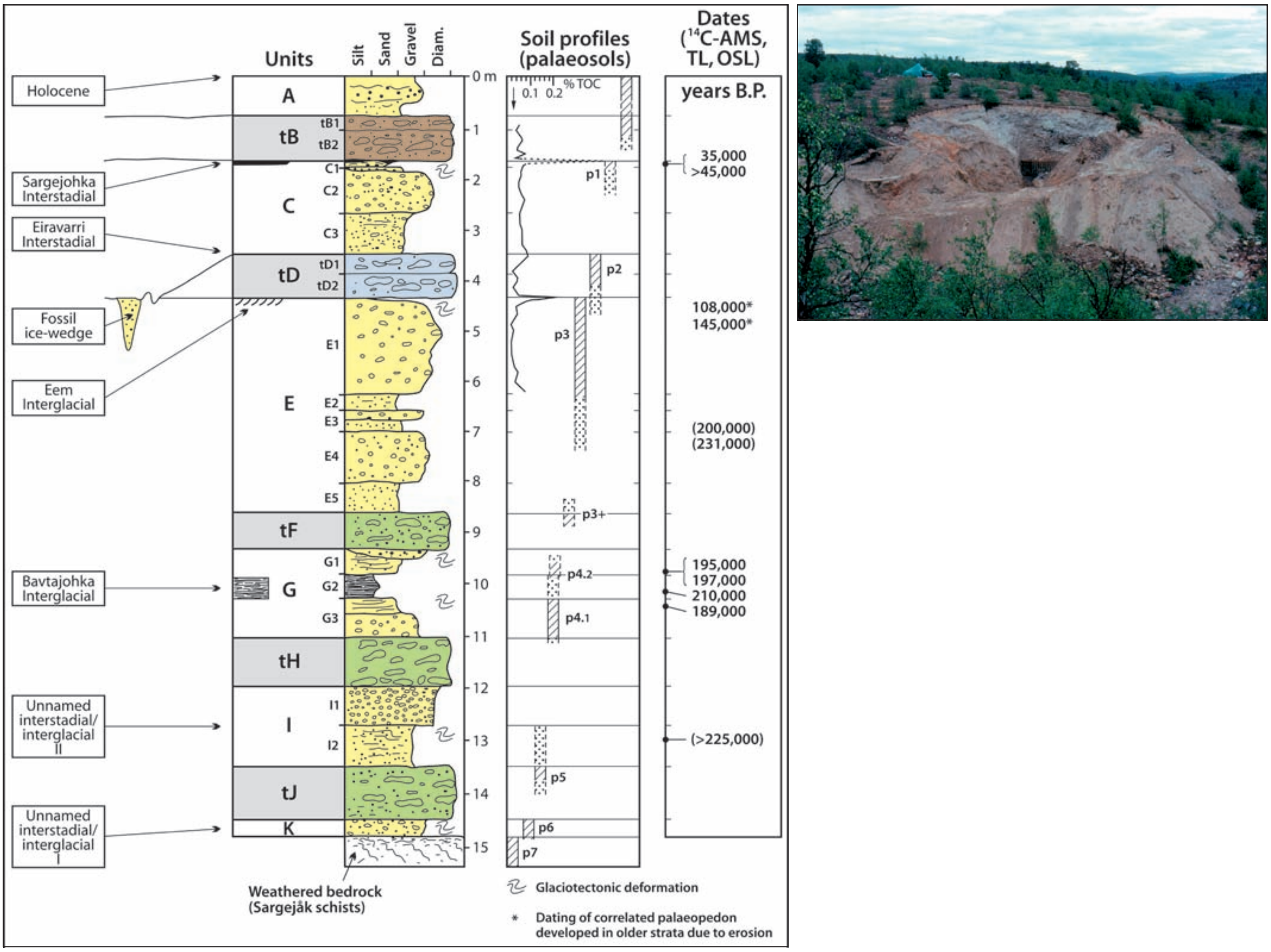


Figure 14. Stratigraphical data from Sargejohka, eastern Finnmarksvidda, modified from Olsen et al. (1996a) and Olsen (1998). (a) Left panel - *stratigraphical column, and (b)* Right panel - *Photograph from excavation at Sargejohka 1989. Dark sediment layers from Bavtajohka interglacial are exposed at c. 10 m depth in the middle of the photo.*

'warm' foraminifera of probably last interglacial (Eemian) age are recorded at some sites in Trøndelag (e.g., Olsen et al. 2001a, 2002), but the only reported in situ sediments of this age is at Skarsvågen on Frøya island, Sør-Trøndelag (Figure 4). This site includes, in addition to Weichselian sediments, both a pre-Eemian, probably Saalian till and marine Eemian beds (Aarseth 1990, 1995).

The Mid-Norwegian shelf. From the Mid-Norwegian shelf a sediment core from the Draugen field includes in its lower part a pre-Weichselian, 70 m–thick succession of till and fine-grained sediments deposited from suspension (Rise et al. 2002, 2005a, The Mid-Norden Project 1999: Map of Quaternary stratigraphy, locality no.104) (Figure 16). At Trænabanken, thick till beds recorded as the Middle and Lower tills, based on data from sediment cores, are assigned Saalian and pre-Saalian ages, respectively (e.g., Dahlgren 2002). Tills and glaciomarine diamictons from the Mid-Norwegian shelf have been correlated with the Naust Formation (glacial debris flows) on the North Sea Fan (e.g., Nygård 2003), and the associated glacier extensions of these units are included in Appendix B, Table B3 (see Substages 9–13, with comments in the west column of maximum ice extension). For more details about the glacial drift on the Mid-Norwegian shelf, we refer to the recent paper by Ottesen et al. (2009).

Northern and central Sweden. Of the 18 reported localities with pre-Weichselian tills in northern and central Sweden only

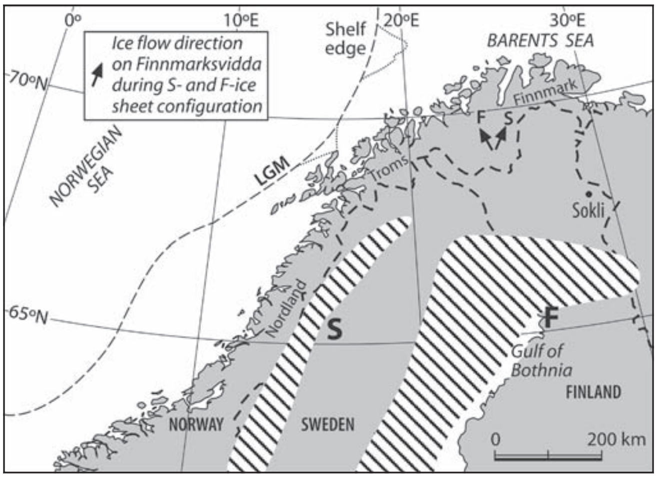


Figure 15. Map of northern Fennsoscandia, with locations of ice-divide zones (shaded) and associated ice movements (arrows) on Finnmarksvidda, modified from Olsen et al. (1996a). F= Fennoscandian ice-sheet configuration, which occurred during maximum glaciations and ice flow towards north-northwest in Finnmark. S= Scandinavian ice-sheet configuration with ice divide along the highest mountains, moderate ice extensions and ice flow towards north-northeast in Finnmark.

two include more than one such till bed. These are Åkåskielas (no.55) and Lainikvare (no.140) from positions halfway and ¾ distance from the Norwegian–Swedish national border to the Bothnian Bay (The Nordkalott Project 1986: Map of Quaternary geology, sheet 4, The Mid-Norden Project 1999: Map of Quaternary stratigraphy) (Appendix B, Figure B3). The

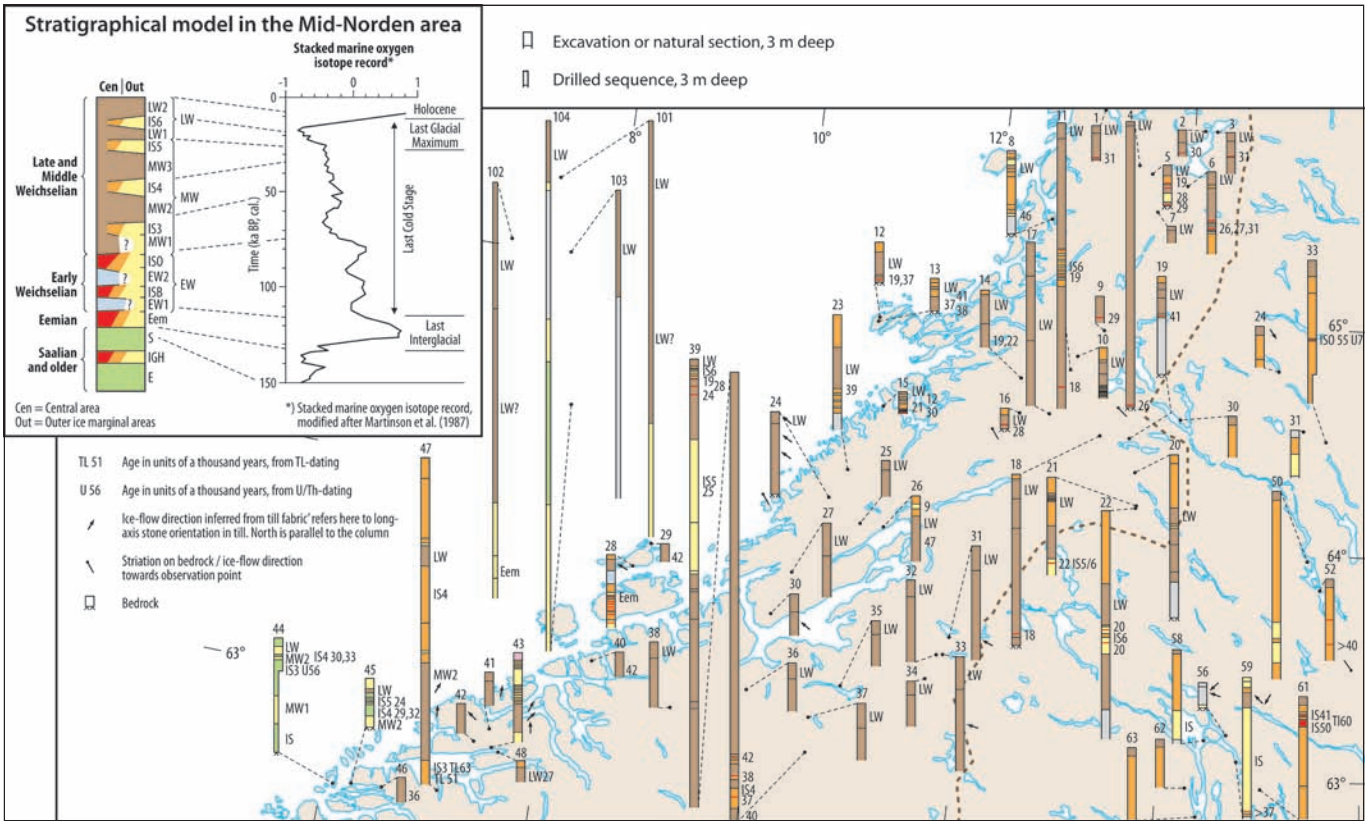


Figure 16. The western part of the Quaternary stratigraphical map of central Fennoscandia (The Mid-Norden Project 1999). Inserted figure: generalised stratigraphical model for the Mid-Norden area. The deposits from the stadials are mainly tills and the colours of these units (green, light blue and brown) in the stratigraphical model have been used similarly in most logs in this paper. Red indicates gyttja, peat, etc., orange= coarse-grained waterlain sediments, and yellow= fine-grained sediments.

occurrence of the oldest till units at these sites implies a glacier extension of at least moderate size of a mountain-centred glacier. Clast fabrics from seven localities with Saalian tills from northern Sweden (The Nordkalott Project 1986: Map of Quaternary geology, sheet 4, nos. 12, 23, 55, 57, 126, 130 and 150) (Appendix B, Figure B3) indicate deposition from a glacier of F-configuration (Figure 15) if all the till beds are from one substage, or more likely deposition both from glaciers of S- and F-configurations if the tills are from more than one substage.

Northern and central Finland. Of the 38 reported localities with pre-Weichselian tills from northern and central Finland only four include more than one such till bed, and only one of these is located in northern Finland (The Nordkalott Project 1986: Map of Quaternary geology, sheet 4, no. 39 Rautuvaara) (Appendix B, Figure B3). The three other of these sites are located less than 140 km from the Gulf of Bothnia in central Finland (The Mid-Norden Project 1999: Map of Quaternary stratigraphy, nos. 16, 36 and 43) (Appendix B, Figure B4). In addition, three sites with glacial striation from pre-Saalian ice flows are reported from the middle and eastern parts of central Finland (nos. 17, 27 and 38) (Appendix B, Figure B4).

The most important site for reconstructing the pre-Weichselian glacier fluctuations and ice flow directions in Fin-

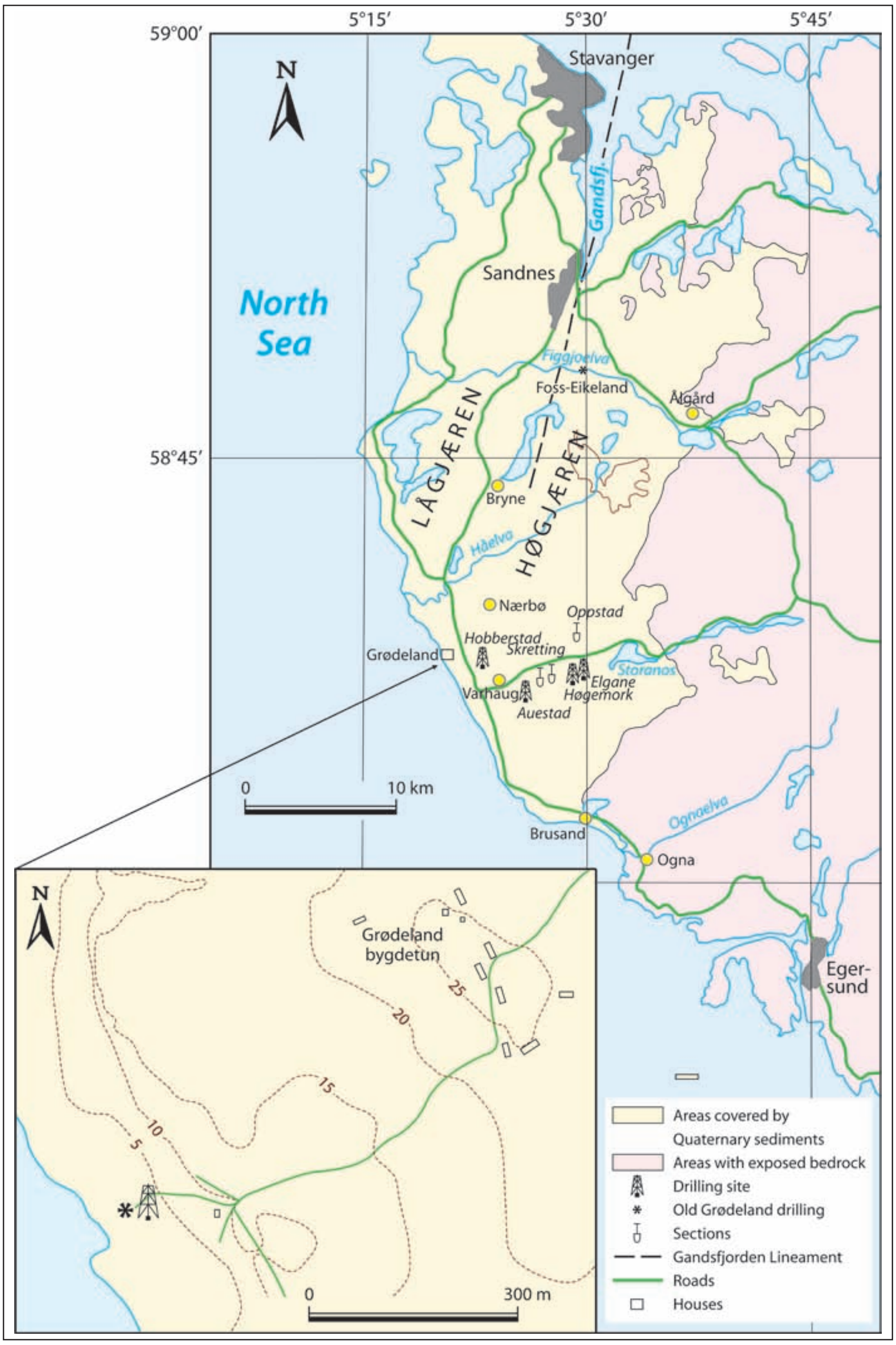


Figure 17. Map of the Jæren area showing the locations of stratigraphical sites (boreholes, sections) referred to in Figures 18, 29 and 31.

land is the Rautuvaara locality (no. 39, Appendix B, Figure B3), near Kolari in northern Finland. It includes five superimposed till beds, three of which are of pre-Weichselian age (Tills 6–4), and this site is a 22 m-high sediment succession that is supposed to be a nice parallel to the Norwegian Sargejohka site mentioned above.

Based on 19 clast fabric analyses, the associated ice-flow directions are inferred to be towards the east to east-southeast for the oldest unit, till bed 6, towards the southeast for till bed 5, and towards the north-northeast (lower part), the east-northeast (middle part) and the south-southeast (upper part) for till bed 4 (Hirvas 1991). Although solifluction may have influenced some of the fabrics, the results altogether suggest associated ice sheets with S-configuration (Figure 15) for till bed 6, F-configuration for till bed 5 (pre-Saalian tills), and both F- (lower and upper parts) and S- (middle part) configurations for till bed 4 (Saalian till).

All the relevant stratigraphical ice-flow directional information from the Rautuvaara locality and the other reported Finnish, Swedish and Norwegian localities with pre-Weichselian tills from northern and central Fennoscandia is compiled in Appendix B (Figure B1) in the accompanying paper by Olsen et al. (2013).

Southern Norway—with additional comments to the North Sea

The oldest reported Quaternary sediments on land in southern Norway are from the Jæren lowland in the southwest, where glacial, interstadial and interglacial deposits with marine fossils, from MIS 10 and upwards are described in boreholes (Janocko et al. 1998, Sejrup et al. 1998, 1999) (Figures 17 and 18). From the coastal area at Fjøsanger, southwest of Bergen (Figure 4), a Saalian till, the Paradis till (MIS 6) is underlying Eemian interglacial marine (MIS 5e) and younger sediments (Mangerud et al. 1981b). Furthermore, at Lillehammer in the inland of southeast Norway a thick till (the Skjellerud till) of pre-Middle Weichselian, probably Saalian age underlies deposits of Middle to Late Weichselian age (Olsen 1985b, 1998).

Data from the Troll core 8903 (Figure 19) (Sejrup et al. 1995) and seismostratigraphical information, also from the Norwegian Channel (Figure 20) (Sejrup et al. 2000, Rise et al. 2004) indicate tills from at least one Saalian (sensu stricto) and four pre-Saalian major ice advances. There seems to be a long nonglacial interval with deposition of a thick marine sediment unit (unit C) between the till bed at *c.* 1.1 million yr BP and the subsequent till that may have a MIS 12 age (Sejrup et al. 1995, Nygård 2003). Recently it has been reported that several glacial erosional horizons occur underneath the base of the previously published Troll core 8903, and the oldest of these may have an age of *c.* 2.7 million years (A. Nygård, pers. comm. 2007). This supports the earlier data from IRD in deep-sea sediments, e.g., from the Vøring Plateau (Bleil 1989, Jansen and Sjøholm 1991), with the implication that glaciers have indeed reached beyond the coastline as early as this.

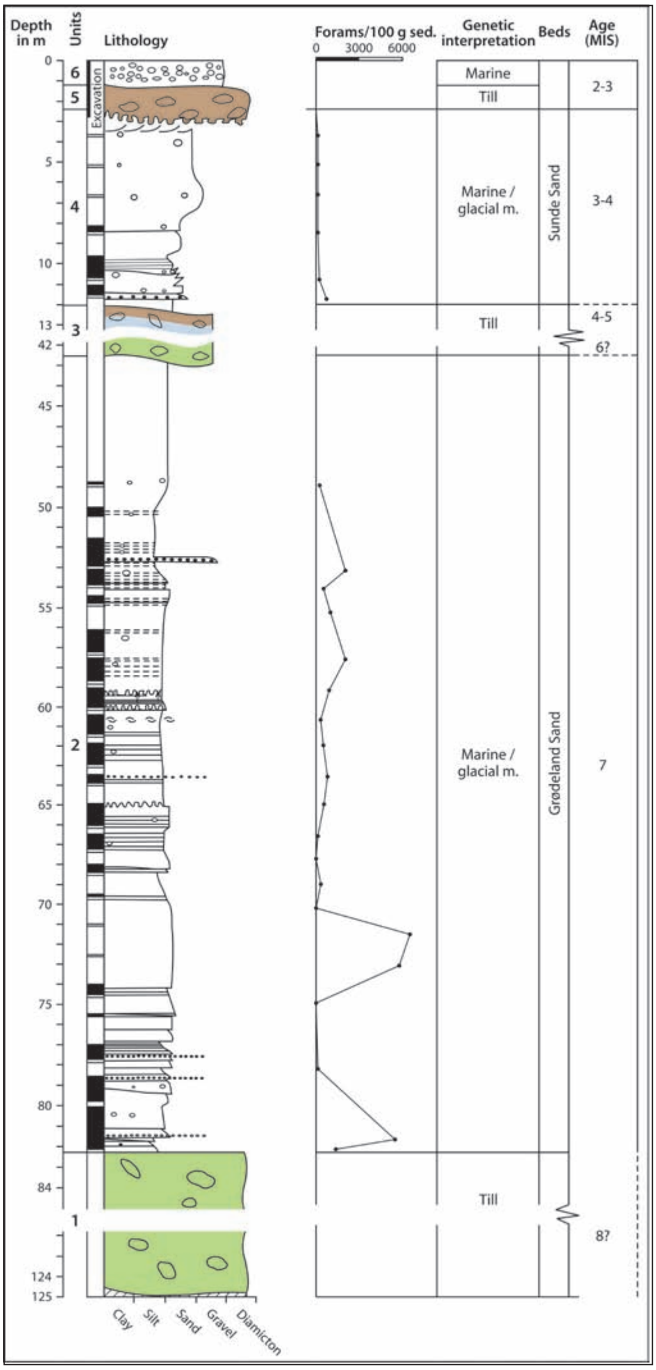


Figure 18. The lithostratigraphy of the Grødeland core, modified from Janocko et al. (1998) and Sejrup et al. (1999). Location of site indicated in Figure 17. Based on amino acid racemisation of marine shells from units 2 (the Grødeland Sand) and 1 (the lowermost till) are supposed to have an age corresponding to MIS 7 and possibly MIS 8, respectively.

Adjacent land areas in the south, southeast and east

Most of the relevant information from the areas considered here are from the compilation of pre-Weichselian glaciations given by Ehlers (1996). During the maximum phases of the Scandinavian Ice Sheet the ice margin reached southwards to the West-European lowlands in the area of the Netherlands–Germany–

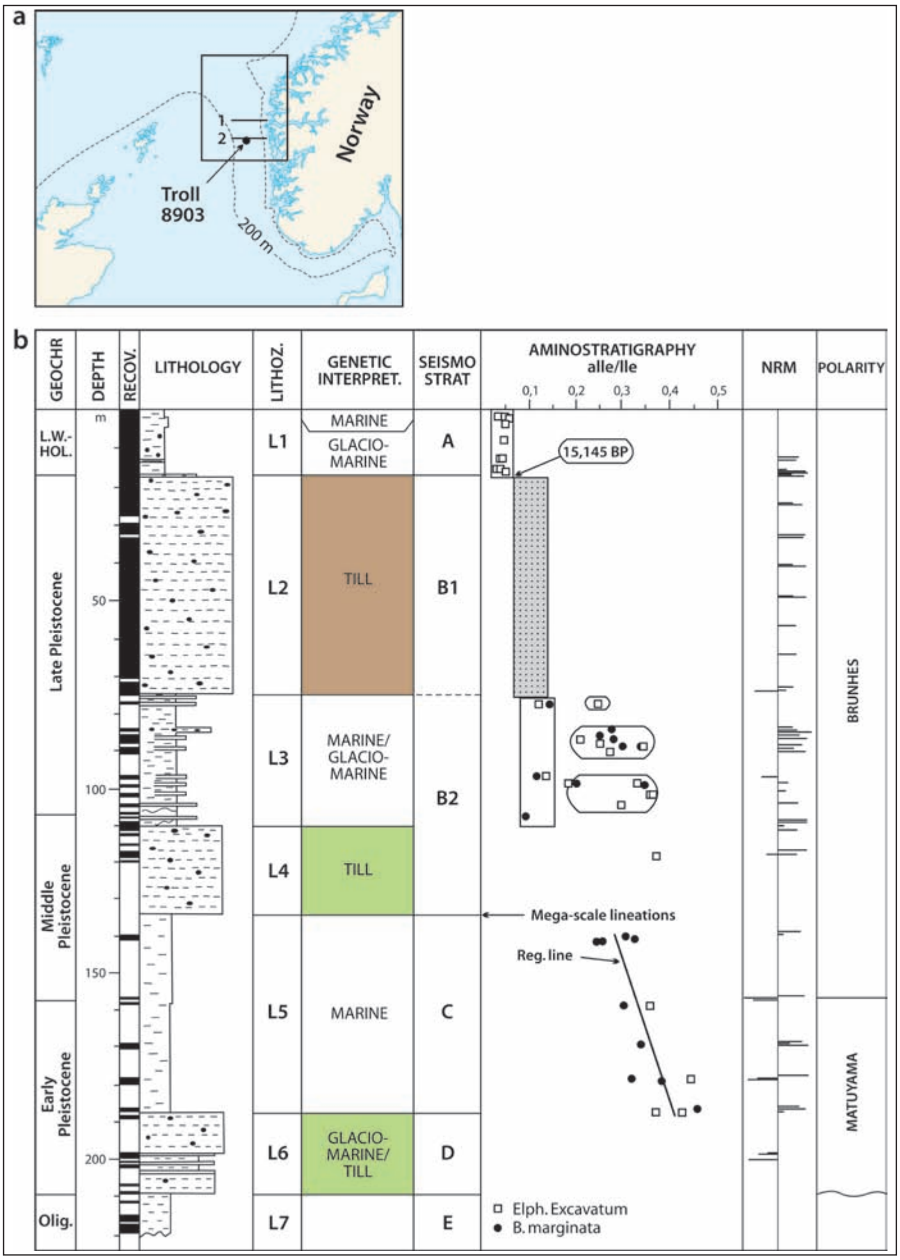


Figure 19. (a) Location map for the Troll core 8903 and the sparker profile lines 1 and 2 referred to in Figures 20a and b, and (b) litho- and chronostratigraphy of the Troll core 8903, and correlation to seismic units indicated in Figure 20. Modified from Sejrup et al. (1995), mainly after Rise et al. (2004).

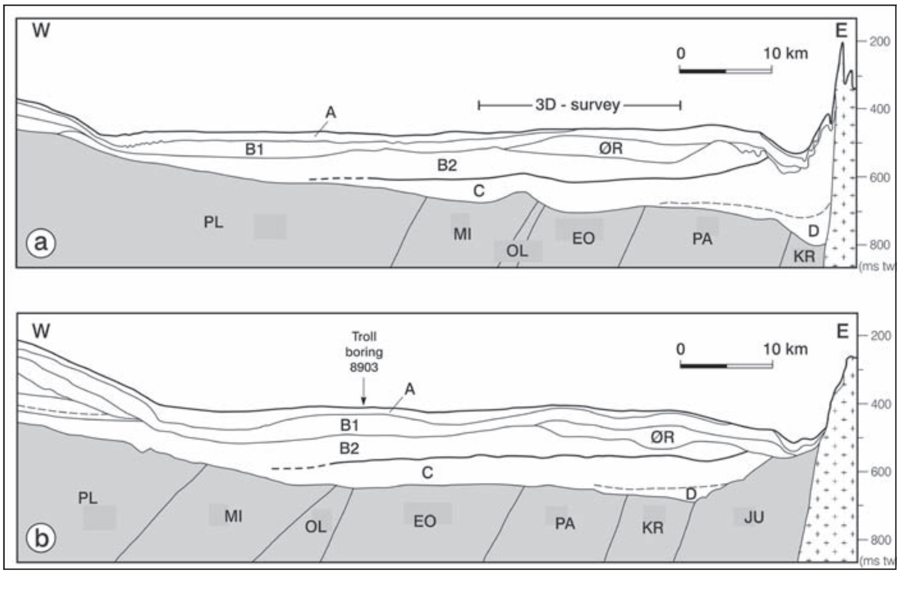


Figure 20. Interpretation of seismostratigraphical cross-section of the Norwegian Channel, (a) along sparker line, profile 1: A77–117, and (b) profile 2: A77–123. After Rise et al. (2004). For location of profiles, see Figure 19a (1= A77–117, and 2= A77–123).

Poland. The glacial deposits from these phases are indirectly dated and correlation to the marine isotope stratigraphy based on the age assignments to the interglacial beds that occur stratigraphically overlying, underlying and/or between these beds (Appendix B, Table B2). Although old glacial deposits have been searched for a long time only the three last glaciations, the Weichselian (MIS 2–5d), the Saalian (MIS 6) and the Elsterian (most likely MIS 12) are documented by glacial deposits/tills in Germany (Ehlers 1996). However, Scandinavian rocks (erratics) that indicate glacial transport are found in older sediments in the Netherlands (Zagwijn 1992).

In the southeast and east (including Russia and Belarussia) the glacial stratigraphical record encompasses tills that are correlated with the Saalian and the Elsterian, and it reaches further back in time and includes also tills from the major Don glaciation (possibly of MIS 16 age, Ehlers 1996) that also had its core area in Scandinavia.

An overview of deposits from the last interglacial (the Eemian)

The Norwegian record of deposits from the Eemian is not large, but still very important as it deals with an important chronostratigraphical marker both on land and offshore (e.g., Mangerud et al. 1981b, Sejrup and Knudsen 1993, Olsen 1998, Hjelstuen et al. 2004, Nygård et al. 2007).

The record includes only some 24 sites on land (Figure 21a and Appendix B, Table B6), but it comprises different formations, such as waterlain sediments (organic and inorganic, marine and terrestrial), palaeosols and speleothems. Several of these formations include palaeoclimatic signatures (as warm as, or warmer than the present) that make them useful for regional correlations. The recorded Eemian sites include six localities from Finnmark, with fluvial/glaciofluvial sediments and/or palaeosols (Olsen et al. 1996a), one from Mid-Norway with gyttja and lacustrine sediments overlain by marine transgression sediments (Skarsvågen, Frøya; Aarseth 1990, 1995), two localities with Eemian marine sediments from southwest Norway (Fjøsanger and Bø; Mangerud et al. 1981b, Andersen et al. 1983), one locality in the inland of southwest Norway with a thin Eemian peat overlain by a Weichselian till (Vinjedalen at Vossestrand; Sindre 1979, Eide and Sindre 1987), at least one locality from Hardangervidda with Eemian lacustrine clay (Hovden; Vorren and Roaldset 1977), and one locality with an Eemian palaeosol developed in a supposed Saalian till at Lillehammer, southeast Norway (Mesna; Olsen 1985b, 1998). In addition, speleothems deposited during the Eemian are reported from caves in Nordland (e.g., Lauritzen 1991, Linge et al. 2001), and several sites with resedimented Eemian organics (e.g., shells and pollen) in Weichselian sediments are reported from different parts of Norway (e.g., Vorren 1972, Vorren et al. 1981, Bergstrøm et al. 1994, Olsen et al. 2001c, Olsen 2002).

The three most cited offshore Eemian sites from the Norwegian shelf are also included in Appendix B (Table B6).

Deposits from major Weichselian substages and intervening ice-retreat phases

An overview of the distribution of Norwegian sites with dated Weichselian-aged sediments (from ice-free intervals) that are intercalated between tills, which have been used as age constrains for the tills, are presented in a series of maps and listed in tables in Appendix B. These include sites from the early Middle Weichselian to Early Weichselian (MIS 4–5d) (Figure 21b and Appendix B, Table B7), the early Middle Weichselian (MIS 3, 44–59 cal ka) (Figure 21c and Appendix B, Table 8), the late Middle Weichselian (MIS 3, 29–44 cal ka) (Figure 21d and Appendix B, Table B9), and the LGM interval (MIS 3/2 and 2, 17–29 cal ka) (Figure 21e and Appendix B, Table B10). The average number of till units that overlie the 'ice-free' sediment units from a particular age interval varies proportionally with age, so that older sediments are overlain by more till units than younger sediments. The average numbers are, listed chronologically (4–3–2–1) from older to younger intervals, 2.65, 1.72, 1.39, and 1.16 (or 1.00, for the LGM) (Appendix B, Tables 7–10). This demonstrates that the *'counting from the top'* principle as described in Appendix A, can be applied and may generally work well for the Weichselian glacial stratigraphy of Norway.

The oldest post-LGM onshore sites with glacial oscillation sediments (till–waterlain sediment–till successions) are listed in Table B11 (Appendix B).

Northern and central Norway—with additional comments to northern and central Sweden and Finland

In northern Fennoscandia, among a total of *c.* 3000 stratigraphical localities more than 175 Finnish, 150 Swedish and 50 Norwegian localities with Early and/or Middle Weichselian till beds have been compiled from various maps (e.g., the Nordkalott Project 1986: Map of Quaternary geology, sheet 4; Olsen 1993). Farther south, in central Fennoscandia, among a total of *c.* 2300 stratigraphical localities more than 47 Finnish localities, 63 Swedish and 48 onshore and several offshore Norwegian localities with Middle Weichselian or older till beds are compiled (e.g., the Mid-Norden Project 1999: Map of Quaternary geology, sheet 3; Dahlgren 2002). Examples with stratigraphical information from some of these and other selected localities are briefly described in the following text.

Northern and central Norway. The Sargejohka, Vuoddasjavri and Vuolgamasjohka sites from Finnmarksvidda, northern Norway (Olsen 1988, 1993b, 1998, Olsen et al. 1996a), include till beds (Gardejohka till) from at least one stadial, and waterlain sediments from at least one interstadial (Eiravarri) during the Early Weichselian. The lower boundary of the Weichselian sediment succession is defined by the subjacent Eemian palaeosol, which is developed as a spectacular podzol at Sargejohka (Figure 22a). Glaciotectonic deformation trending east in the waterlain sediments from the last interglacial (Eemian) at Vuolgamasjohka suggests that the first Weichselian ice

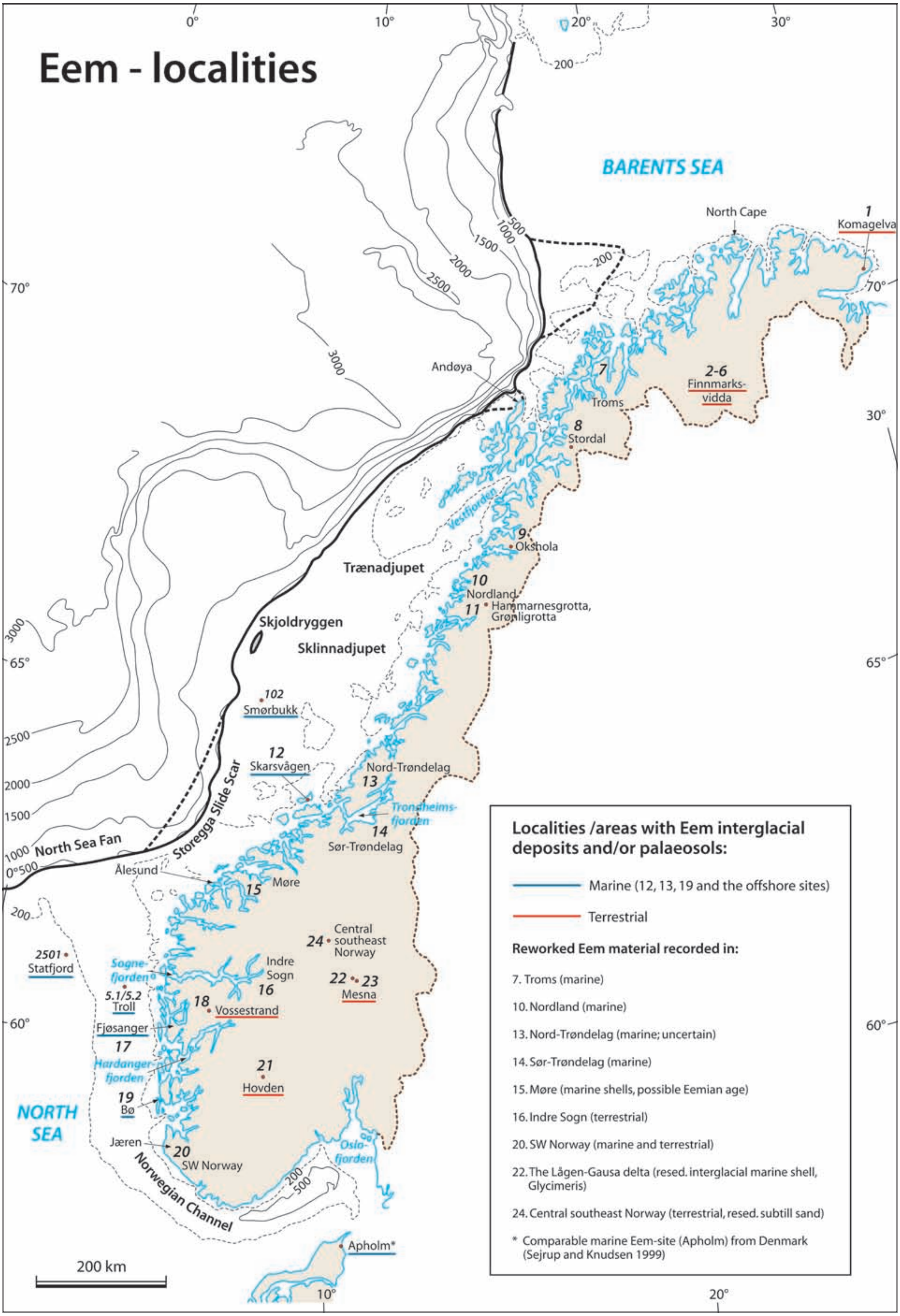


Figure 21. (a) Eem sites (same as Olsen et al., 2013, fig. 6). (b) Early Middle Weichselian to Early Weichselian sites (MIS 4–5d). (c) Early Middle Weichselian sites (MIS 3, 44–59 cal ka). (d) Late Middle Weichselian sites (MIS 3, 29–44 cal ka). (e) LGM interval sites (MIS 3/2 and 2, 17–29 cal ka).

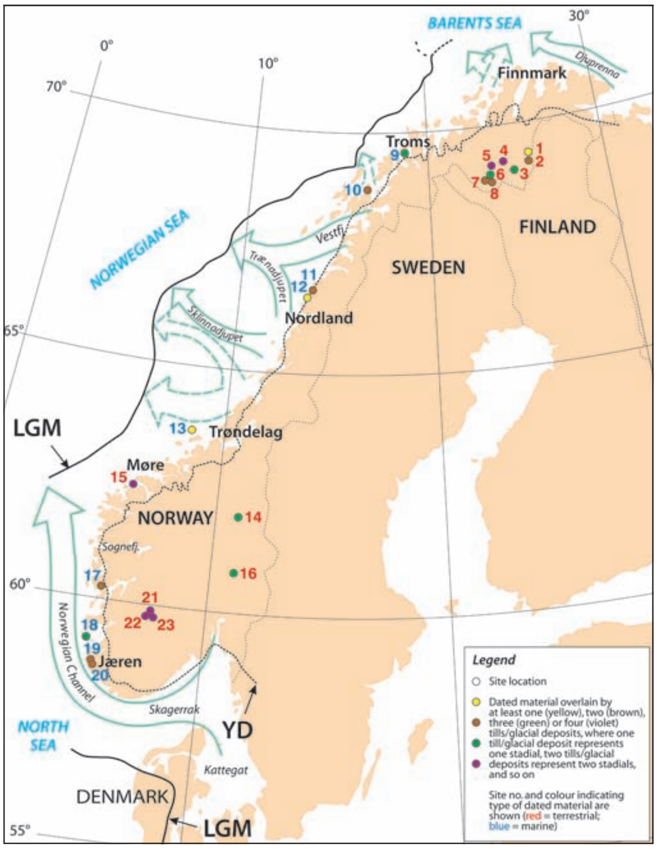


Figure 21b: Early Middle Weichselian to Early Weichselian (MIS 4–5d) (60–115 ka).

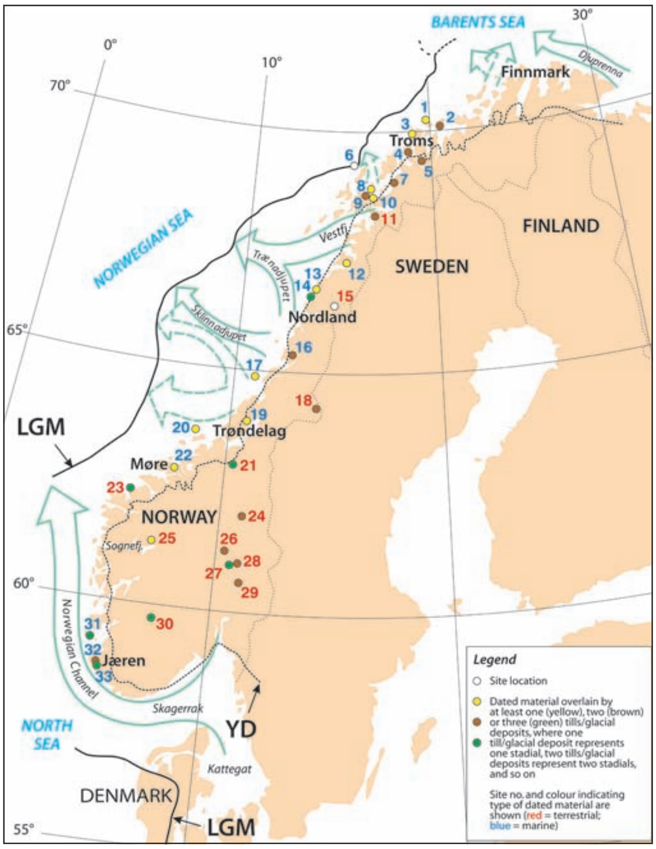


Figure 21c: Early Middle Weichselian (MIS 3) (44–59 cal ka).

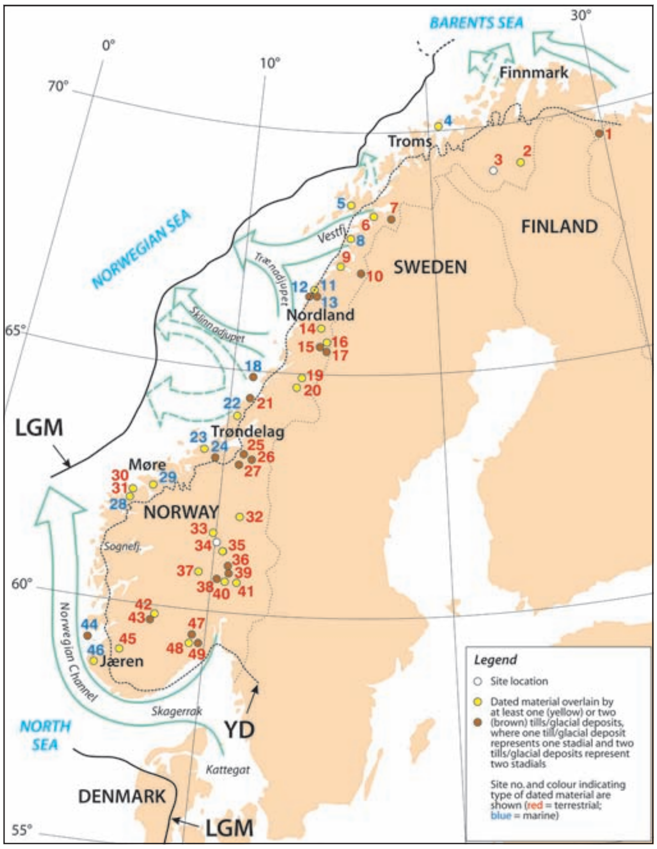


Figure 21d: Late Middle Weichselian (MIS 3) (29–44 cal ka).

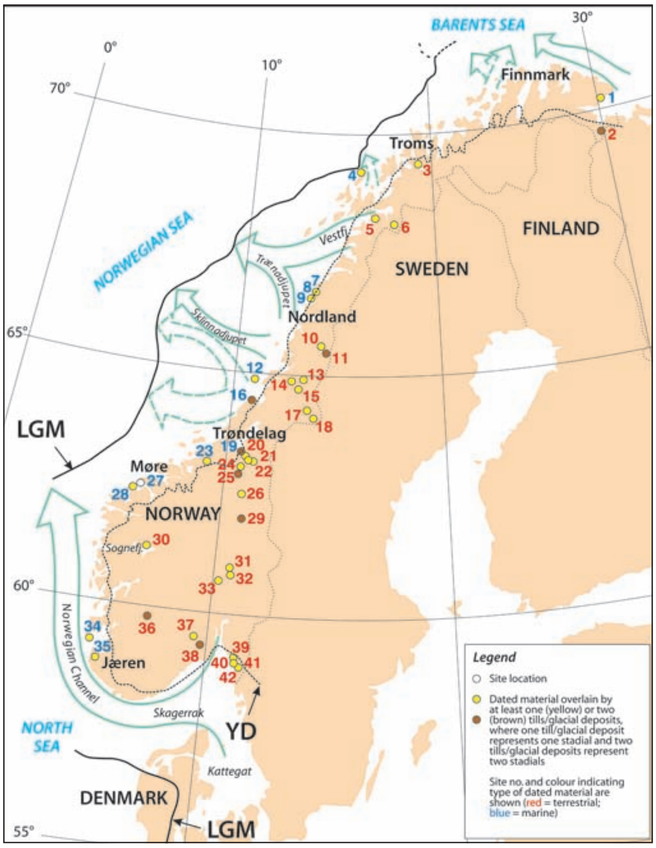


Figure 21e: Late Weichselian (MIS 2), LGM interval c. 15–25 ^{14}C ka (i.e., 17–29 cal ka).

flow came from the mountains in the west. No recorded till is associated with this ice flow, which indicates that the initial Weichselian glaciation may have occurred without deposition of till on Finnmarksvidda (Olsen et al. 1996a). Alternatively, the associated till was eroded during subsequent ice flow(s). All three sites also include tills (Vuoddasjavri till) from at least one stadial, and waterlain sediments from at least one interstadial (Sargejohka) of Middle Weichselian age.

At Leirelva and Komagelva in northeastern Finnmark (Figure 22b), inferred Eemian and older well-sorted sand is overlain by glaciofluvial sediments and till from the Late Weichselian (Olsen et al. 1996b).

At Leirhola, Arnøya, northern Troms, till-covered glaciomarine deposits up to *c.* 8 m a.s.l. indicate that the glacier margin was close to the site a short time before 34–41 cal ka (i.e., around 44 cal ka), advanced beyond the site with deposition of a till bed shortly after *c.* 34 cal ka, retreated and readvanced over the site with additional till deposition shortly after 31 cal ka (Andreassen et al. 1985). A new excavation in the year 2000, *c.* 40 m from the former site revealed a sediment succession including shore deposits and marine-transgression sediments overlain by till indicating that the sea level fluctuated around 14 m a.s.l. in a period sometime between 34 cal ka and the last major LGM advance (Figure 23, Olsen 2010). Due to the considerable glacioisostatic depression, and despite the existence of a globally very low sea level (120–130 m b.s.l.) at that time, a relative sea level as low as this is not likely to represent a phase with rapid ice retreat *c.* 24 cal ka, i.e., during the Andøya–Trofors interstadial between the LGM I and II advances. A comparison with the last deglaciation history suggests that the relative sea level during such conditions would be expected to be at least 20–30 m a.s.l., considered the near maximum glacioisostatic depression at this time. It is much more likely that the 14 m sea level occurred around 31 cal ka, a time when the global sea level was *c.* 50 m lower than the present (e.g., Shackleton 1987) and with the local glacioisostatic depression of moderate size, at the end of an interstadial or after the retreat of a medium-sized glacier expansion and ice-volume event.

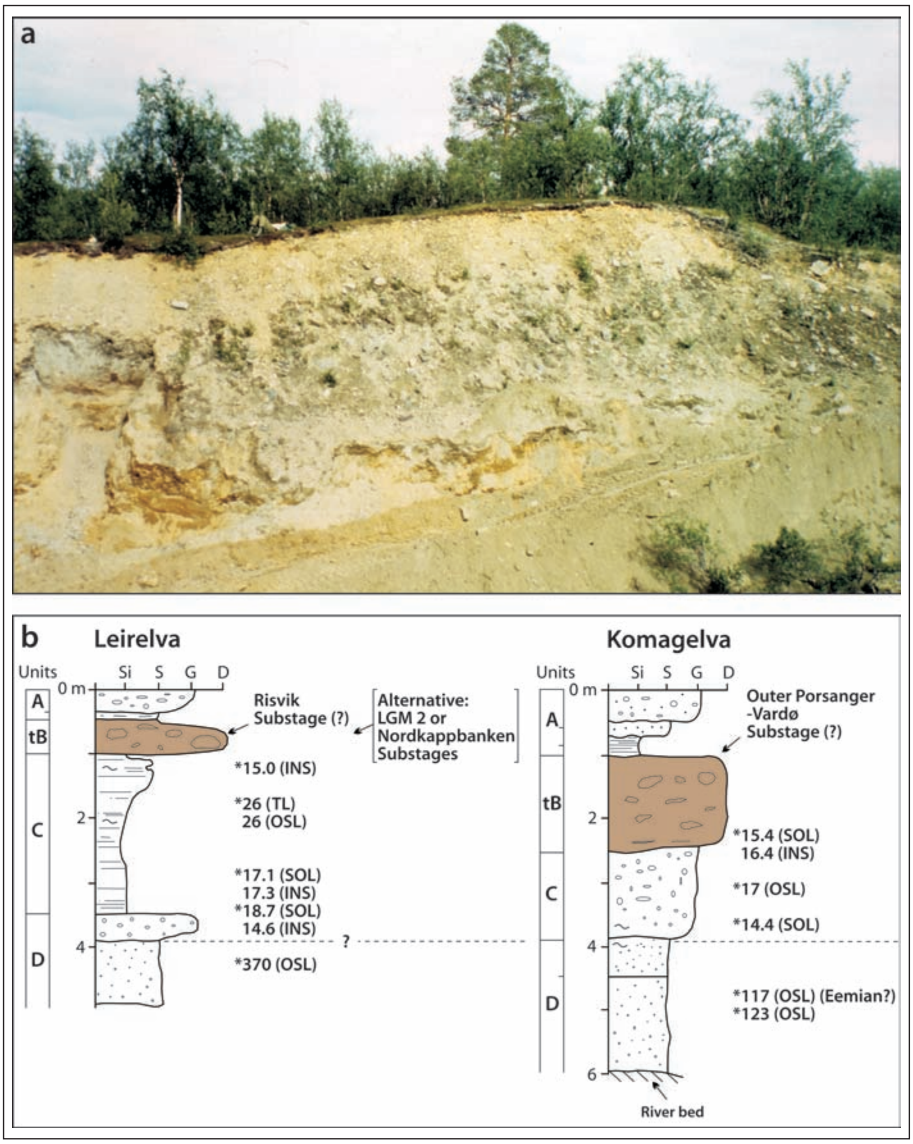
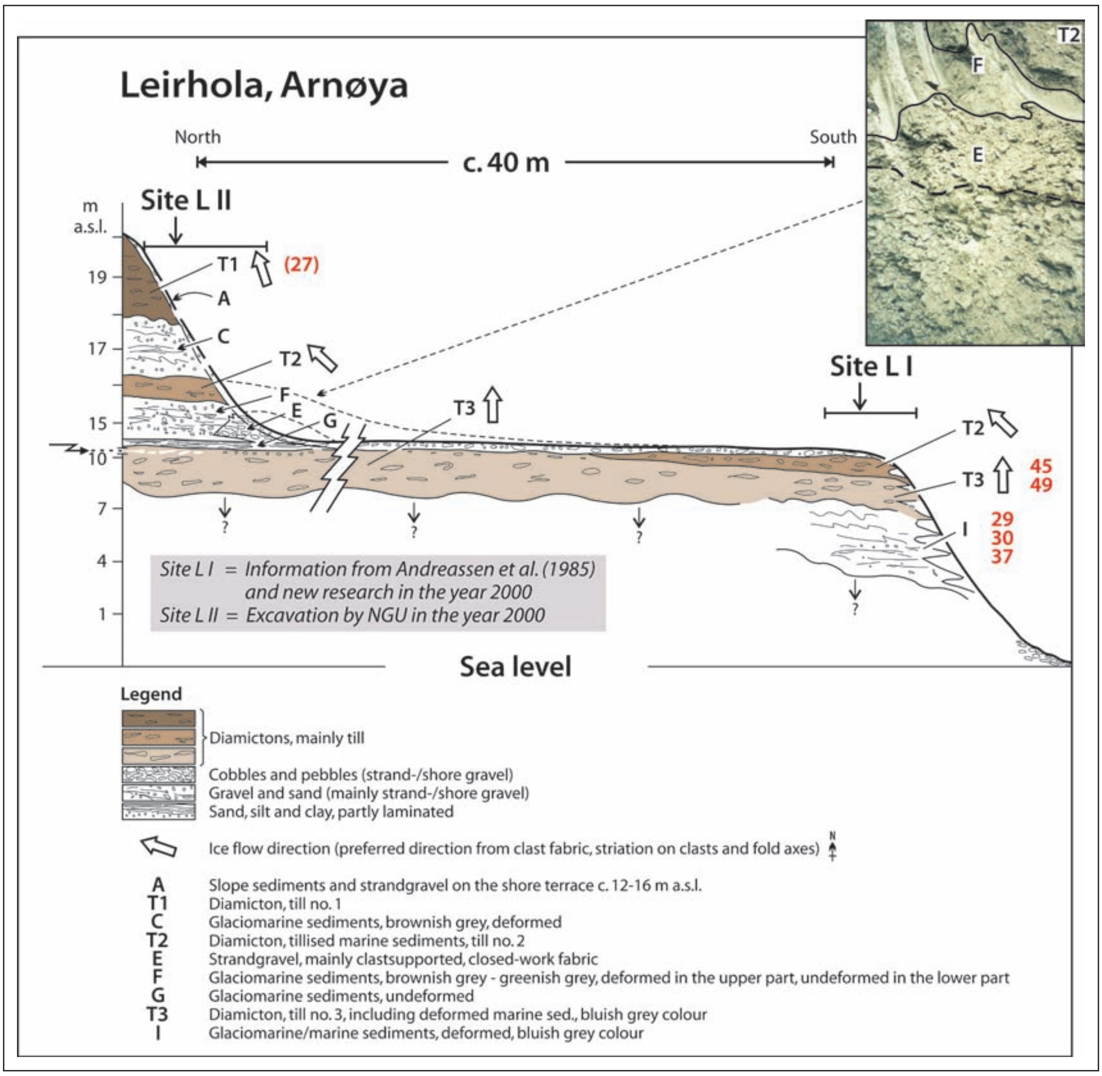


Figure 22: (a) Upper panel – *The strongly cryoturbated Eemian palaeosol, in subtill position and 3–5 m below ground surface at Sargejohka on eastern Finnmarksvidda, North Norway. After Olsen et al. (1996a). (b)* Lower panel – *Correlation of the lithostratigraphies from the inferred Eemian and Weichselian sites Leirelva and Komagelva, Varanger Peninsula, northeast Finnmark, North Norway. Mainly after Olsen et al. (1996a). For location of sites, see Figure 4.*

Figure 23. Generalised section connecting the Middle and Late Weichselian sites LI and LII from Leirhola, at Arnøya (island), North Norway. For location of Arnøya, see Figure 4. Inset photograph: Strandgravel (unit E), between glaciomarine sediments (units F and G), which indicates a distinct lowstand between higher relative sea-levels.



^{14}C dates, from younger to older stratigraphical position, age in ka (red numbers) in Figure 23:

^{14}C - dating (T-3507, 27400 ± $^{1500}_{1200}$ yr BP, Astarte sp. / Mya truncata) from adjacent loc. (Lauksundet), after Andreassen et al. (1985)

^{14}C - dating (TUa-3624, 44755 ± $^{1745}_{1435}$, Arctica islandica / Mya truncata) from T3, Site II

^{14}C - dating (TUa-3626, 48635 ± $^{2595}_{1960}$, Arctica islandica / Mya truncata) from T3, Site I

^{14}C - dating (T-4020, 29000 ± $^{4200}_{2700}$, Chlamys and Portlandia arctica / lenticula; T-4021, 30200 ± $^{4100}_{2700}$, div. Shell sp.; T-3509, c. 37500 yr BP, Mya / Astarte / Arctica islandica)

Shell fragments of age around 44 cal ka in tills are recorded from a number of the islands along the coast of Troms and Nordland. Examples of these are Arnøya, Vanna, Kvaløya, Grytøya, Hinnøya, Langøya and Åmøya, but also localities on the mainland show such finds, e.g., in Balsfjorden, Salangen, Sagfjorden, Salten, Skogreina, Glomfjorden, Bjærangsfjorden and Velfjorden. All these data together indicate an ice advance, possibly around 44 cal ka, that reached beyond the coastline in the west. Some of these and other relevant localities will be briefly described below.

At Slettaelva on Kvaløya, northern Troms, the sediment succession starting at the base on bedrock includes a till from a local glacier trending eastwards, i.e., in the opposite direction compared to the subsequent Fennoscandian ice sheet (Vorren et al. 1981). The till is overlain by laminated lacustrine sediments, which were deposited in a tundra environment and in a pond caused by local damming conditions or dammed by a moraine produced during the advance of the Fennoscandian ice sheet, possibly sometime before 46 cal ka, and perhaps during MIS 4 (Figure 24). On top of these sediments lies a till which is divided in three subunits that each may represent an ice advance over the area. The lower two of these contain shell fragments that are dated to *c.* 46 cal ka and estimated, based on amino acid diagenesis, to represent shells from three different intervals, *c.* 44, 77 and 136 ka (cal), respectively (Vorren et al. 1981). Consequently, three regional glacier advances of which the oldest possibly is *c.* 44 cal ka, may have reached beyond this site, and one earlier regional advance, probably also of Weichselian age, may have reached to the site.

At Løksebotn in Salangen, southern Troms (Figure 25) (Bergstrøm et al. 2005b) four till beds separated and underlain with sand and gravel are recorded. Most units are shell bearing, and most shells are resedimented fragments of *Mya truncata* and *Arctica islandica*, except for those from the sediments representing the last deglaciation, which are whole, small shells of *Portlandia arctica* that is a subsurface sediment feeder and should therefore, if possible, be avoided as a precise dating object (e.g., Mangerud et al. 2006). The two oldest tills (T4 and T3) are from the LGM interval or older. The ice movements in this area suggest that the resedimented shells from the lowermost sand and gravel (which itself is located up to 22 m a.s.l.) derive from higher ground along the fjord, which indicate original sedimentation during glacioisostatic depression. The

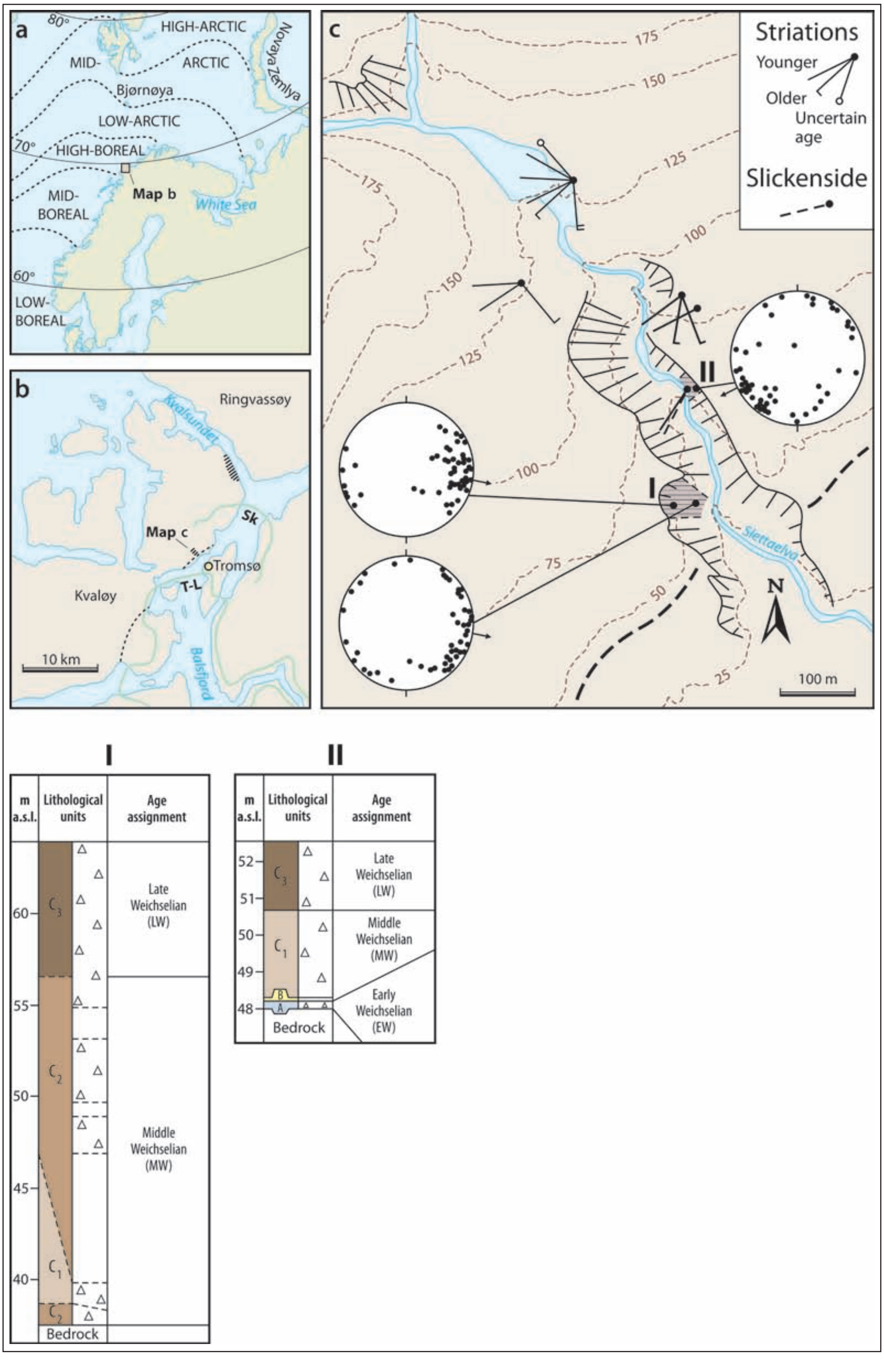


Figure 24. Location maps (a, b, c) of the Middle and Late Weichselian site Slettaelva, North Norway, of which the lithostratigraphy is indicated in logs in the lower panel. Modified from Vorren et al. (1981).

resedimented shells from this locality (Figure 25) are dated to at least a Middle Weichselian age, which indicate that a thick ice covered the ground in the vicinity at, or just before this time, probably around 60 or 44–49 cal ka (the closest known relevant ice-growth phases).

At Storelva, on Grytøya (Figure 4), southern Troms, shallow-marine sediments between till beds *c.* 125 m a.s.l., with shells dated to around 44 cal ka, indicate that the ice was in the vicinity and may have reached beyond Grytøya shortly after this time. It is possible that the lower till also derives from the Weichselian, e.g., MIS 4 (Olsen and Grøsfjeld 1999, Olsen et al. 2001c).

A glacier has probably reached Bleik on Andøya around 44 cal ka, because shells from glaciomarine sediments on the low

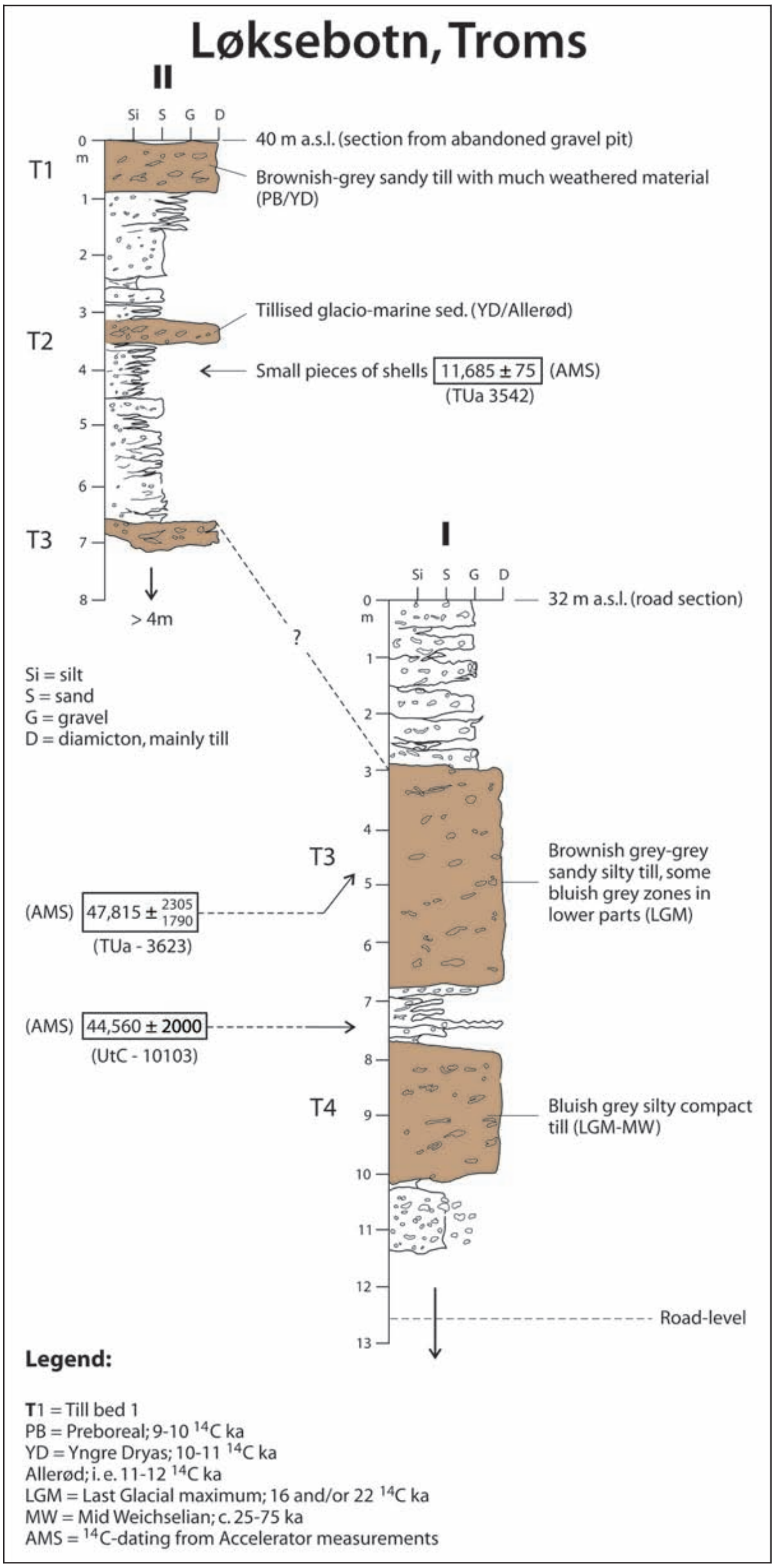


Figure 25. Lithostratigraphy of Middle and Late Weichselian deposits from Løksebotn, Troms, North Norway. Modified from Bergstrøm et al. (2005b). For location of site, see Figure 4.

ground distally to the Bleik area are dated to this age (Møller et al. 1992). At Trenyken, one of the outermost small islands in the Lofoten area, Nordland, a ^{14}C-dating of shell fragments from sediments washed over a moraine into a cave *c.* 19 m a.s.l., gave an age of 33,560 +/- 1100 ^{14}C yr (*c.* 37.5 cal ka) (Ua–2016, Møller et al. 1992). The moraine was deposited at the mouth of the cave, possibly around 44 cal ka. However, redeposition from older strata is an alternative hypothesis, which suggests that the moraine may by younger than 37.5 cal ka. In any case, with a 44 cal ka or LGM age of the moraine, the high relative sea level (more than 19 m a.s.l.) during marine inwash of the shell-bearing sediments is a clear indication of a phase with a considerable glacioisostatic depression far offshore in this western shelf area.

Shell dates from a complex ice-marginal formation at Skogreina, central Nordland, suggest that the ice front may have been there around 44 cal ka (Olsen et al. 2001c, Olsen 2002). At Hundkjerka, a sediment-filled cave in coastal southern Nordland, a ^{14}C-dating of a resedimented shell fragment from marine silt and sand overlying a diamicton with many weathered clasts and underlying diamict material, till and marine sediments from the last deglaciation gave an age of *c.* 50 cal ka (Figure 26) (Olsen et al. 2001c). The level of the cave mouth, *c.* 40 m a.s.l., indicates clearly that the marine sedimentation occurred during a phase of considerable glacioisostatic depression, similar to the first 1000–2000 years of the Holocene. This could have happened just after the retreat of the 44 cal ka ice advance, which consequently must have been considerable in this area. However, the data from this site do not rule out the possibility

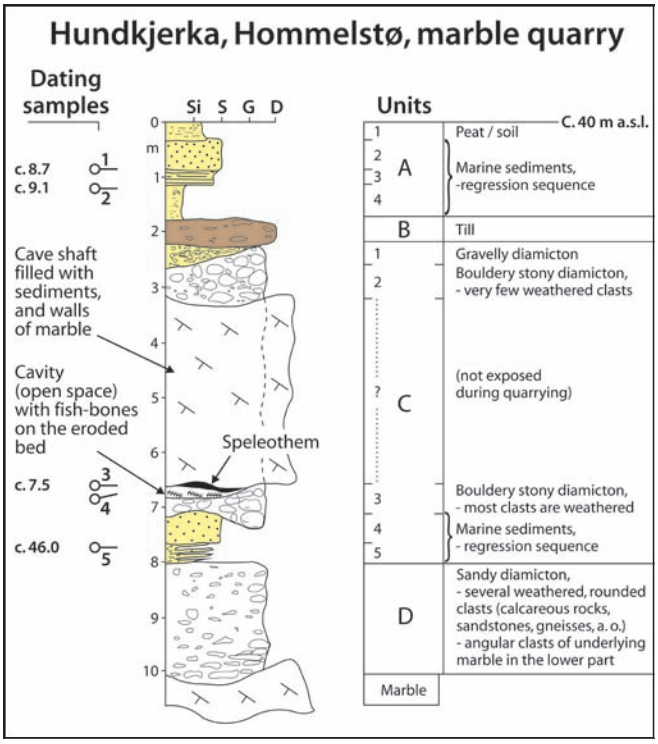


Figure 26. Lithostratigraphy of the sediment succession in the Hundkjerka cave, Nordland, North Norway. Ages in ^{14}C ka BP. After Olsen et al. (2001c). For location of site, see Figure 4.

of an older age (dating close to the upper limit/range of the method) or a younger age (resedimentation) of this ice phase, for example *c.* 34 cal ka.

Plant fragments from sediments redeposited in till at Lierne, in the inland of Nord-Trøndelag, and shells in till at Vikna, on the outer coast of Nord-Trøndelag are both dated to *c.* 44 cal ka, which give a maximum age for the subsequent regional ice advance (Olsen et al. 2001c).

Glacial stratigraphies with intercalated waterlain sediments at Selbu, in the inland of Sør-Trøndelag, represent a highly dynamic late Middle Weichselian to LGM interval ice sheet (Figure 27a). OSL dates at 21–22 cal ka from waterlain sediments correlated with the Trofors interstadial (from the middle of the LGM interval) have recently been reported from the Langsmoen site and support the earlier published ^{14}C-ages of plant remains from subtill sediments at the neighbouring Flora site (Olsen et al. 2001a, Johnsen et al. 2012).

At Grytdal, in Gauldalen south of Trondheim, Sør-Trøndelag, ^{14}C-dated plant remains in tills and intercalated waterlain sediments indicate several Middle to Late Weichselian ice-margin fluctuations in that area, and that the oldest of these ice advances occurred *c.* 44 cal ka (Figure 27b, Olsen et al. 2001c).

At Svellingen, on Frøya (Figure 4), Sør-Trøndelag, shells in till are dated to *c.* 46 cal ka, which may indicate an ice advance *c.* 44 cal ka reaching beyond the Frøya Island (Aarseth 1990). At Skarsvågen, also on Frøya, Sør-Trøndelag, Eemian marine shell-rich sediments and terrestrial sediments (gyttja) are overlain by Early Weichselian marine-transgression sediments followed by regression sediments and till, which is covered by a Middle to Late Weichselian till on top (Aarseth 1990, 1995). The marine sediments overlying the Eemian gyttja derive from a deglaciation, which may either follow a MIS 5d, 5b or a MIS 4 glaciation, perhaps with the latter as the most likely alternative. In that case the overlying till is likely to be of MIS 3 (*c.* 44 cal ka) age, whereas the younger till bed may be either of MIS 3 (*c.* 34 cal ka) or MIS 2 age. However, the age problem for these deposits is not yet solved (I. Aarseth, pers. comm. 2004).

Ice-dammed sediments in the caves Skjonghelleren and Hamnsundhelleren at Møre (Figure 4) indicate that glacier advances reached beyond the coastline of western Norway at least four times, separated by ice-free conditions/interstadials during the Weichselian (Larsen et al. 1987, Valen et al. 1996). The first ice advance crossed these caves at *c.* 60 ka, the next *c.* 44 cal ka, the following *c.* 32 cal ka, and the last after 28.5 cal ka. Geomagnetic excursions (Laschamp at *c.* 44 cal ka and Lake Mungo/Mono Lake at *c.* 32 cal ka) that are recorded in the sediments confirm the age model for the two ice advances in the middle, and therefore indirectly also the other advances (see also review by Mangerud et al. 2003).

Luminescence dates around 100 ka of glaciogenic sandur sediments at Godøya, near Ålesund, Møre, indicate the ice-margin position, or retreat of a relatively large glacier, possibly of MIS 5d age (Jungner et al. 1989, Mangerud 1991a, b, c). However, a younger age, e.g., MIS 5b, or even a Middle Weichselian age of this event, as originally suggested by Land-

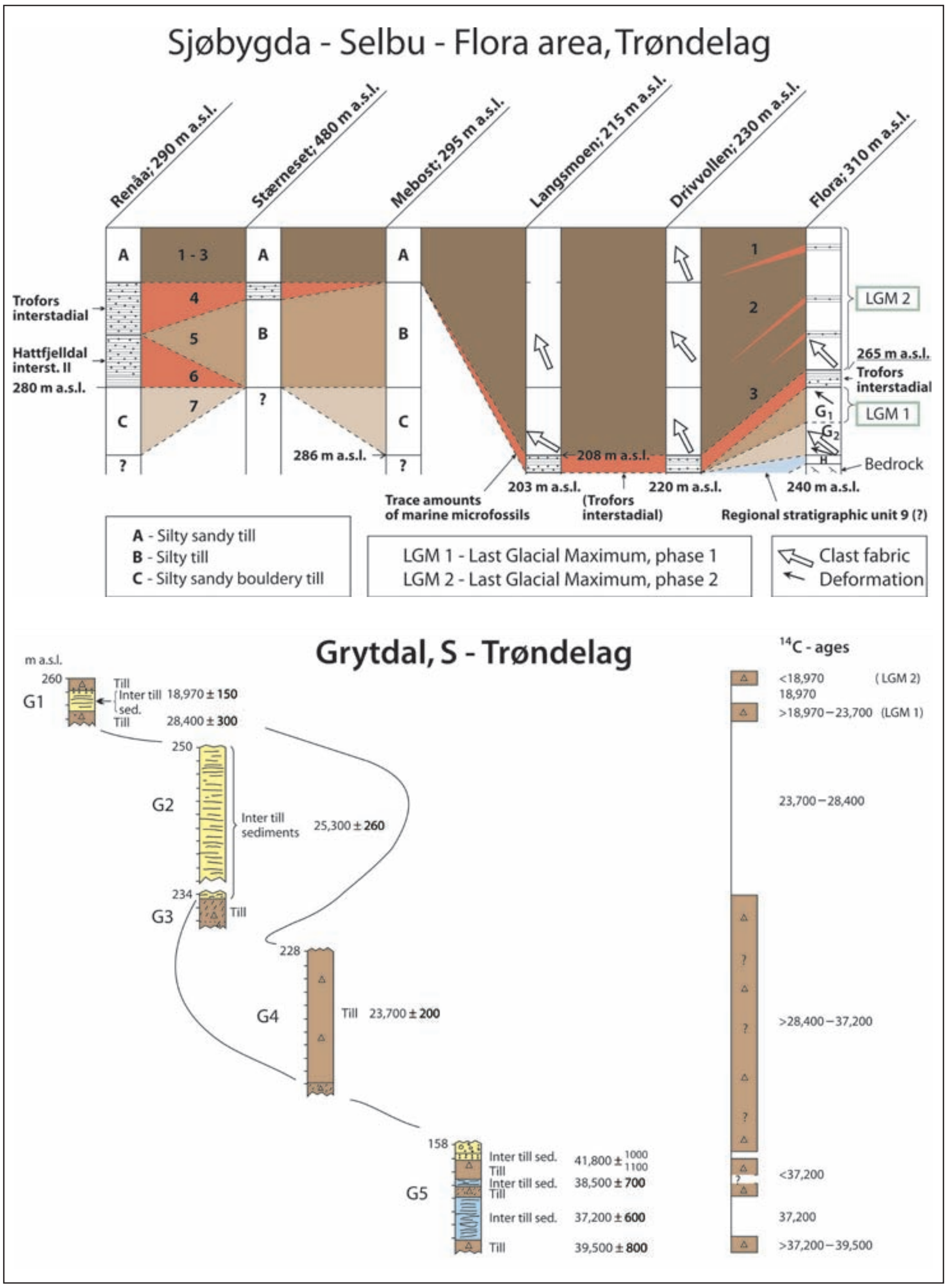


Figure 27. (a) Upper panel - *Glacial stratigraphies from the inland of central Norway, east of Trondheim. Modified from Olsen et al. (2002). OSL dates at 21–22 cal ka from Trofors interstadial-correlative sediments at the Langsmoen site have been reported by Johnsen et al. (2012).*

(b) Lower panel - *Correlation of lithostratigraphies from sections G1–G5 in Middle and Late Weichselian sediments along a steep local road in Grytdal (main valley: Gauldalen), central Norway. Modified from Olsen et al. (2001a). For location of sites, see Appendix B, Table B1–B, and Figure B1.*

vik and Mangerud (1985) and Landvik and Hamborg (1987), should still be considered.

Northern and central Sweden. Till beds correlated with the Finnish Early Weichselian Till 3 are recorded at many sites in northeastern Sweden (The Nordkalott Project 1986: Map of Quaternary geology, sheet 4), and are so far assumed to be present at two sites in central Sweden (The Mid-Norden Project 1999: Map of Quaternary geology, sheet 3, sites no. 43 Stornipan and 47 Hoting). Till 3 or other Early Weichselian tills may well be represented at other sites in this region as the first till unit overlying sediments from the Jämland interstadial (Lundqvist 1969), but this is not properly demonstrated yet. However, tills from the Middle Weichselian (e.g., correlated with the Finnish Till 2) are commonly represented at many sites, both in the northern and central Sweden.

Northern and central Finland. All sites where Till 3 is recorded are located proximal to the Pudasjärvi moraine in northern/central Finland (Figure 4, Sutinen 1984), and consequently, no sites with till from the Early Weichselian is recorded in central Finland (The Mid-Norden Project 1999). The age of Till 3 is uncertain, but the Nordkalott project data (Hirvas et al. 1988) combined with the new information from the Sokli area (Helmens et al. 2000, 2007) suggest that the till may be subdivided in an older unit (which may correlate with MIS 5d), not represented at Sokli, and a younger unit from a more extensive advance, correlated with MIS 5b.

Another important glacial boundary in northern Finland is represented with the Lapland moraines, which are thought to derive from an Early or Middle Weichselian advance or retreat phase.

The Middle Weichselian Till 2, which followed the Peräpohjola interstadial (Hirvas et al. 1988, Hirvas 1991), is more commonly represented at many sites, both in the northern and central Finland (The Nordkalott Project 1986, The Mid-Norden Project 1999).

Southern Norway—including some data from adjacent areas in the south, southeast and east

Southern Norway. In southern Norway, coastal sites as Fjøsanger, Bø on Karmøy and several sites at Jæren are important sources of information with regard to glacial deposition during the Early to Middle Weichselian (e.g., compilation by Mangerud 1991a, b, c, 2004) (Figures 4 and 28a, b). It has been shown that Early Weichselian stratigraphies including glacial deposits exist in the coastal areas, which indicate glacial advance at least once, followed by an ice-retreat/interstadial phase and another ice advance that crossed the coastline, deposited a till (Bønes till) in southwestern Norway and reached the Norwegian Channel at least once during the Early Weichselian. Middle Weichselian marine sediments, which are located at highly elevated sites at Jæren (>200 m a.s.l.) (e.g., Andersen et al. 1987, Larsen et al. 2000, Raunholm et al. 2004), indicate deposition during a state of considerable glacioisostatic depression and therefore indirectly that a big ice must have existed there and/or in the Norwegian Channel shortly before. Consequently, a Middle Weichselian glacier extension has reached the coast at Jæren (Figure 29), and probably also the Norwegian Channel (where an ice stream may have existed) at least once (Sejrup et al. 1998, 2000, Larsen et al. 2000). The Bø site on Karmøy and the Foss–Eikeland site at Jæren include also important Late Weichselian stratigraphies (Figures 30–31), that imply considerable fluctuations of the LGM-interval ice sheet in southwestern Norway.

In the inland areas of southwest Norway, in Vinjedalen at

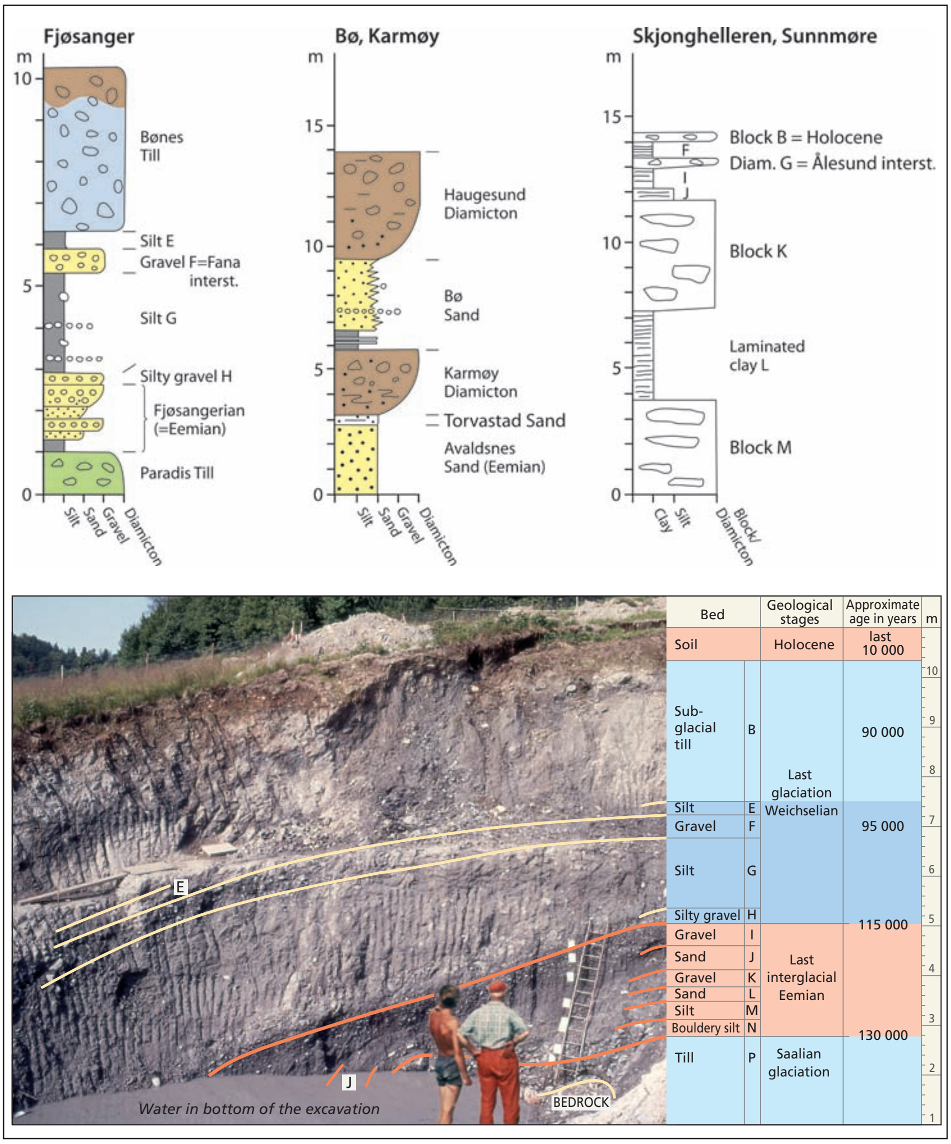
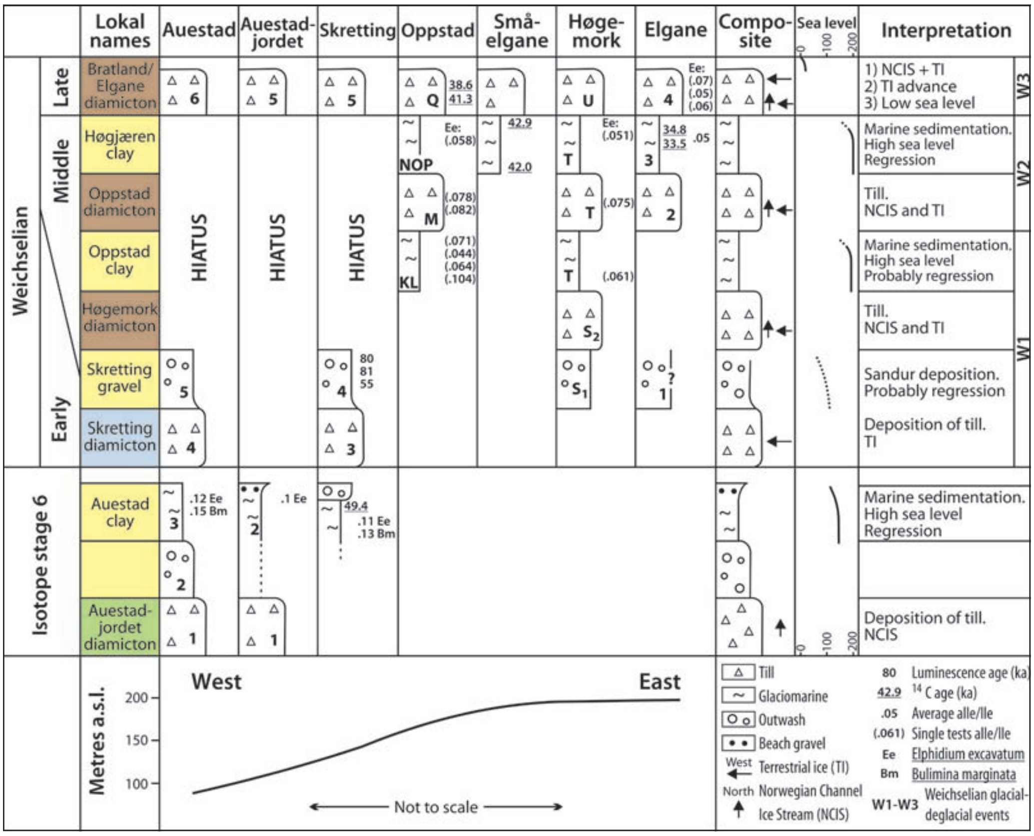


Figure 28. (a) Upper panel – *Eemian and Weichselian sediments in the three stratigraphical sites Fjøsanger, Bø and Skjonghelleren (cave). After Mangerud (1991). (b)* Lower panel – *Photograph from the Fjøsanger site. Note that the colours in (b) of the units are not standardised as they are in the other illustrations in this paper. Modified from Vorren and Mangerud (2006). For location of sites, see Figure 4.*

Figure 29. Saalian and Weichselian stratigraphies from sediments at sites from Jæren, southwest Norway. After Larsen et al. (2000). For location of sites, see Figure 17.



Vossestrand (Figure 4) a 2 cm-thick highly compressed Eemian peat is overlain by a Weichselian till (Sindre 1979). In Brumunddalen, southeast Norway till beds (units K, L, P and Q) from the Early Weichselian are recorded underlying the Brumunddalen interstadial beds at Dalseng (Helle 1978, Helle et al. 1981) (Figure 32a), and the interstadial peat was highly compressed at 59 cal ka, possibly by an ice advance, as indicated by a U/Th-dating of this age (S.-E. Lauritzen, pers. comm. 1991). At Mesna, Lillehammer,

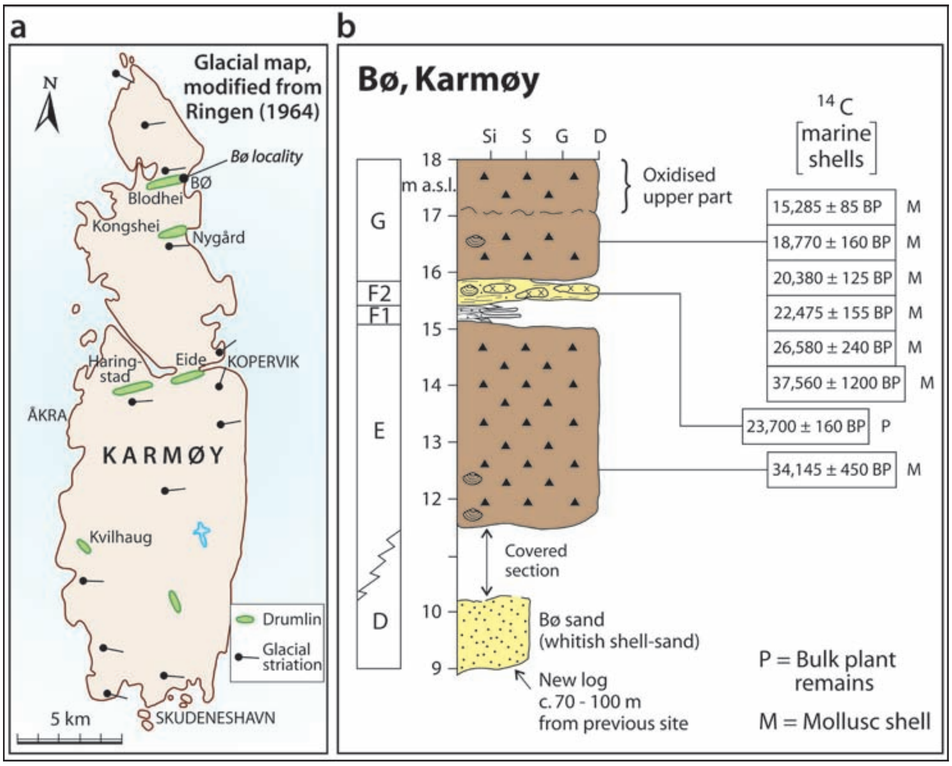


Figure 30. (a) Location map, and (b) lithostratigraphy of the upper part of the sediments in the drumlin at Bø, Karmøy. After Olsen and Bergstrøm (2007). The deeper parts of the sediments at this site (Bø, in Figure 28a) have been studied by Andersen et al. (1983) and Sejrup (1987). The new units E, F1, F2 and G indicated here, represent a subdivision of the previously described uppermost unit, the Haugesund diamicton.

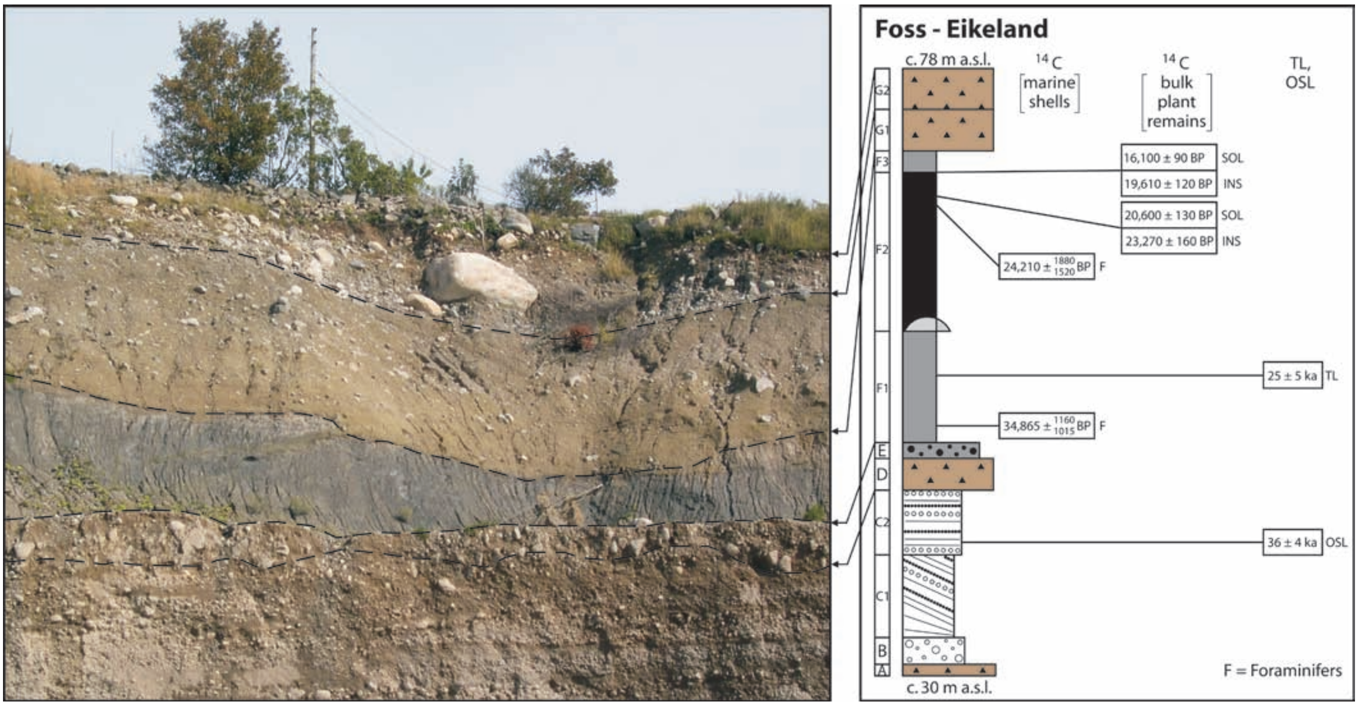


Figure 31. Photograph and lithostratigraphy from the gravel pit in the Middle and Late Weichselian sediments at Foss-Eikeland, Jæren. Lithostratigraphy modified from Raunholm et al. (2003). For location of site, see Figure 17. Photograph by B. Bergstrøm 2002.

southeastern Norway, a till bed (the Mesna till) of age framed between 40 and 44–48 cal ka is recorded (Figures 32a, b and 33b, unit A), and at Rokoberget farther southeast, highly elevated Middle Weichselian sediments with marine microfossils indicate a considerable glacioisostatic depression around 44 cal ka, similar to that recorded for the Jæren area (Rokoengen et al. 1993a, Olsen and Grøsfjeld 1999, Olsen et al. 2001c, Larsen et al. 2000).

At Hardangervidda (Figures 33 and 34), glacial stratigraphies from nine localities reveal a history that according to Vorren and Roaldset (1977) and Vorren (1979) includes deposits from the Eemian (Hovden thermomer), from one stadial (Hovden kryomer) and one interstadial (Førnes thermomer) supposedly from the Early Weichselian, and from two Middle to Late Weichselian stadials separated by one interstadial (all representing the complex Førnes kryomer). Correlations to other areas were made on the basis of pollen content and inferred palaeovegetation and palaeoclimate, but with no absolute dates available. Recently, OSL dates of subtill sediments from Hovden (at Møsvatn), which is one of the sites described by Vorren and Roaldset (1977), and from Grytkilvika (at Mårvatn), previously described by Vorren (1979), and from two additional sites in this area have been reported by Haug (2005). He found that these dates indicate a Middle to Late Weichselian age of all the studied subtill interstadial sediments, including the Hovden sand that Vorren and Roaldset (1977) and Vorren (1979) correlated with the Førnes thermomer, which they assumed was of Early Weichselian age. Based on the OSL dates from Hardangervidda, Haug (2005) suggests a correlation with the three or four Middle to Late Weichselian interstadials, pre-Hattfjelldal (*c.* 50 cal ka), Hattfjelldal 1 (*c.* 37 cal ka), and Hattfjelldal 2 (*c.* 30 cal ka) or Trofors (*c.* 22.5 cal ka), that have been described and reviewed from Norwegian inland areas by Olsen (1997) and Olsen et al. (2001b).

Southern Sweden–Denmark–Poland. At Dösebacka–Ellesbo (Figure 4) in the vicinity of Gothenburg, southwestern Sweden, three Weichselian tills are separated with intercalated pollen-bearing interstadial sediments (Hillefors 1974). The oldest interstadial is probably from the Early Weichselian and the younger one from the Middle Weichselian. The oldest till is therefore possibly of either MIS 5d or 5b age, most likely the latter, whereas the middle till is likely to derive from MIS 4, or possibly MIS 3, since glaciers during both these stages are supposed to have reached southwards beyond the Gothenburg area. The youngest till bed is from the LGM interval (MIS 2).

At Sjötorp, a few km southwest of Skara (Figure 4) in southwestern Sweden, glaciotectonised waterlain sediments have been TL and OSL dated to 43–45 cal ka (Ronnert et al. 1992). The deformation movement was directed from the W–WNW, which indicates that an ice movement from the west may have reached this area and tectonised the sediments around *c.* 44 cal ka, or possibly during a younger ice advance, e.g., *c.* 34 or 25 cal ka. Houmark-Nielsen and Kjær (2003) discussed in their palaeogeographic reconstructions an E-trending ice movement in this area and favoured a 27–29 cal ka age of this.

As reviewed by, e.g., Lundqvist (1992), only till beds from the Late Weichselian have been found overlying the Eemian interglacial sediments at the classical site Stenberget, in the

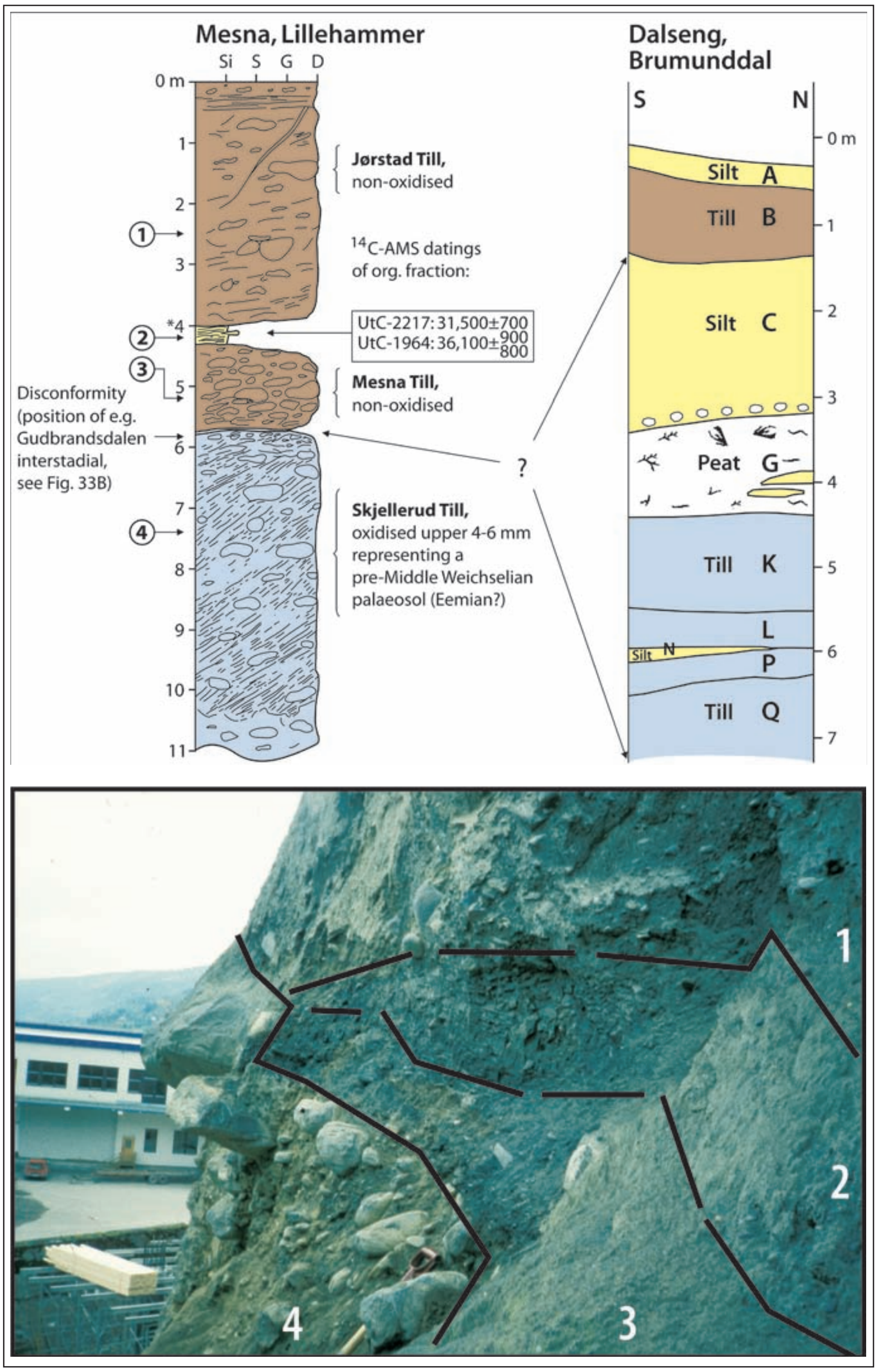


Figure 32. (a) Upper panel – *Lithostratigraphy of inferred Saalian to Weichselian deposits at Mesna, Lillehammer. To the right, a several metres thick Early, Middle and Late Weichselian sediment sequence from Dalseng (units from Helle 1978) is supposed to correlate with the position of the disconformity between units 3 and 4 at Mesna. Modified from Olsen (1985b, 1998). (b)* Lower panel – *Photograph from the Mesna section (L. Olsen 1990). Stratigraphical units 1–4 are indicated. For location of sites, see Figure 33a.*

southwesternmost part of Sweden. Similar data are recorded also from the Alnarp valley, east of Stenberget (Miller 1977). Only Late Weichselian tills overlie the Eemian interglacial beds there too.

Glacial deposits which are TL-dated to 60 cal ka indicate that the first ice advance from the Fennoscandian Weichselian ice sheet may have reached Poland during MIS 4. Furthermore, stratigraphical data comprising tills, glaciofluvial sediments and glaciotectonics indicate that ice advances from the east, from the Baltic basin reached the islands in southeastern Denmark both during MIS 4 and 3 (see review by Mangerud 2004).

Southern Finland–northwest Russia–the Baltic states. A MIS 5b ice may have reached and deposited till at the Kola Peninsula, in northwest Russia (Lundqvist 1992, Mangerud 2004). Except for that, no glaciers seem to have covered southern Finland, northwest Russia, Estonia, Latvia or Lithuania during the Early Weichselian. From a compilation of the Weichselian glacial history based on the glacial stratigraphy for the Baltic states made by Satkunas and Grigiené (1996), the first Fennoscandian Weichselian glacial advance reached this area during MIS 4, as indicated by till beds of this age recorded in Estonia and possibly in a narrow belt in adjacent areas of Latvia. According to these

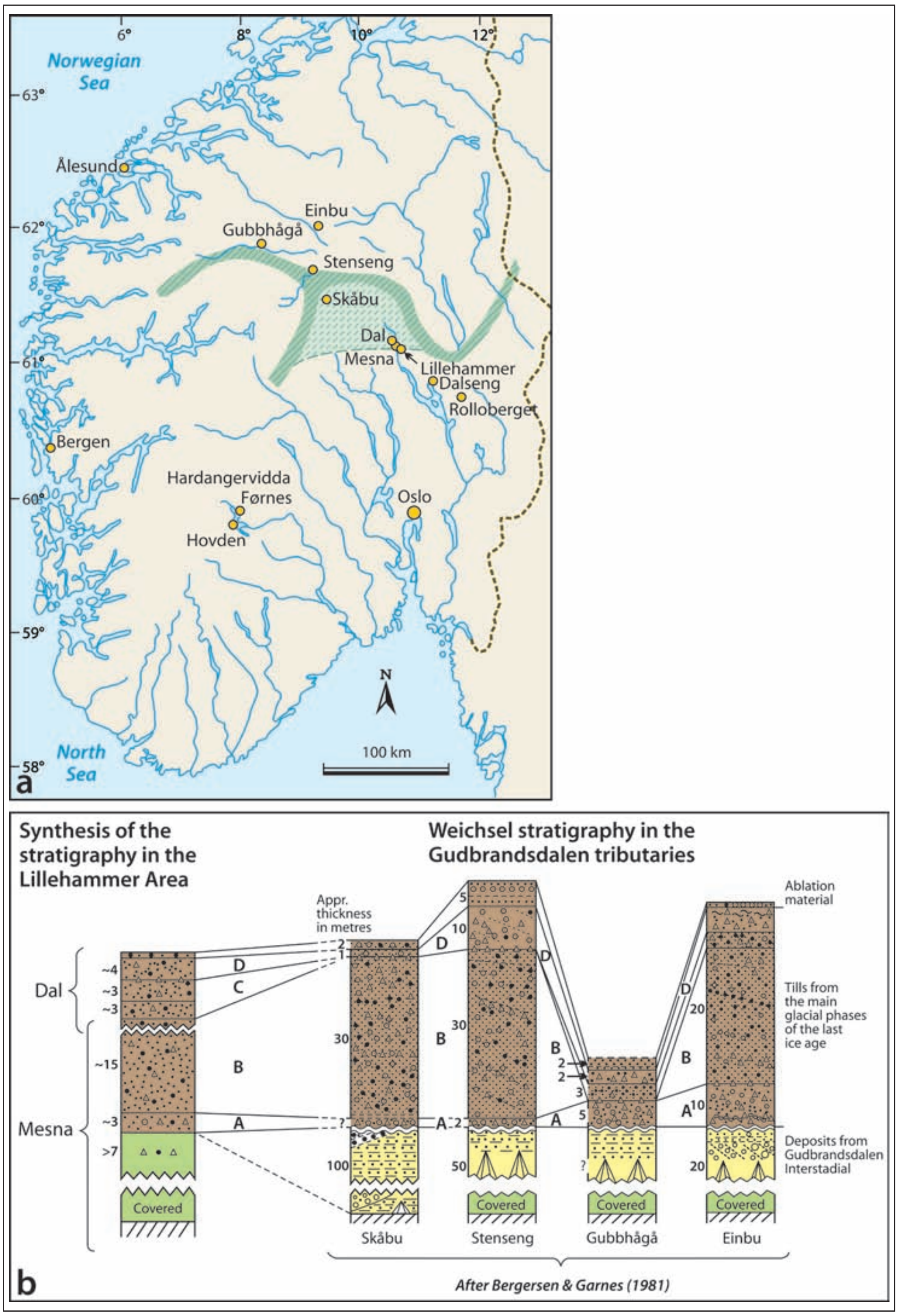


Figure 33. (a) Location map, with late LGM-interval ice-divide zone (shaded), and (b) correlation between lithostratigraphies from the Lillehammer and Gudbrandsdalen area, southeast central Norway. After Olsen (1985b).

authors, the glacial advance during MIS 3 may have reached to Estonia, but not as far as to Latvia or Lithuania.

Offshore areas in the north, northwest, west and southwest

New information on glacier dynamics and ice-front deposits on the shelf has come from the discovery and interpretation of submarine fans at the mouth of glacial troughs ('trough-mouth fans'), and the existence of distinct ice streams across the shelf during glaciations. A review of these is given by Vorren et al. (1998). More recently it has been shown, through detailed morphological mapping of the sea floor, that there have been a number of ice streams across the Norwegian shelf (Figure 6) (Ottesen et al. 2001, 2005) and in the southwest Barents Sea area (e.g., Andreassen et al. 2008). There are numerous large, parallel glacial lineations (megaflutes and megascale lineations) produced during combined glacial accumulation–erosion processes in the troughs, whereas features diagnostic of glacial flow are absent from the shallow ground. If such features ever existed in the shallow parts, then subsequent iceberg ploughing may have disturbed and deformed them beyond recognition.

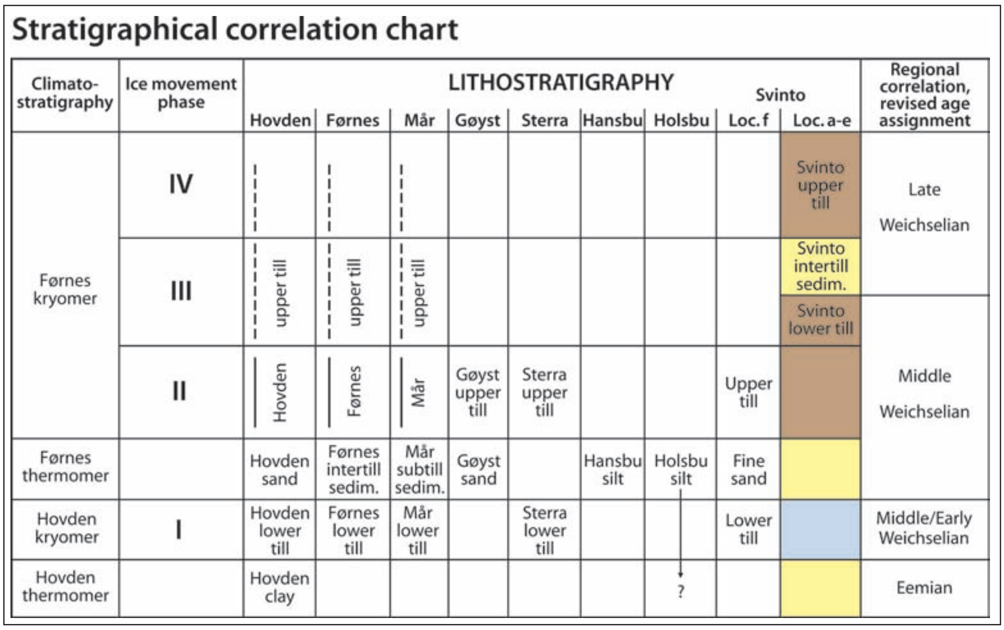


Figure 34. Correlation chart of Late Pleistocene stratigraphies from Hardangervidda, central southern Norway. Modified from Vorren (1979), with revised age assignments based on data from Haug (2005). For location of the Hovden and Førnes sites, see Figure 33a. All other sites in the panel are from the same area.

The interpretation is that there was only slow moving ice on the shallows. This and other relevant data, also from the shelf area, have been reviewed by, e.g., Mangerud (2004).

It is uncertain how many times and how far out on the shelf the western Fennoscandian ice margin reached during the Early/Middle Weichselian. However, both offshore data as well as indirectly also data from onshore positions (Jæren) indicate that the ice sheet grew big enough to produce an ice stream in the Norwegian Channel at least twice in this period (Larsen et al. 2000, Nygård 2003). It seems like most of the authors favour a MIS 4–3 age of these events (Larsen et al. 2000, Nygård 2003, Mangerud 2004). However, the more recent and presumably most precise result from seismic data and coring (MD99–2283) from the North Sea Fan suggests glaciogenic debris flow and corresponding distal laminated sediments from one major pre-Late Weichselian, possibly Early Weichselian ice-sheet expansion (with maximum MIS 5b age, i.e., about 90 ka), and in that case there is no indication of a major Middle Weichselian ice advance (Nygård et al. 2007, and A. Nygård, pers. comm. 2007).

Deposits from the present interglacial (the Holocene)

The erosional and depositional history of the Holocene includes many good examples of the overall and changing environment during an interglacial. However, as glacial deposits are in focus in this paper, and since sediments of the Holocene result from mainly nonglacial processes and represent significant parts of the modern landscape, this period is further reviewed in the accompanying paper that is focused on Quaternary landscape development (Fredin et al., 2013).

Summary and conclusions

An overview of glacial offshore and onshore deposits in Norway is given here (Figures 1–3), with additional examples from selected offshore areas and from several onshore locations from all regions of Norway. The location of sites with interglacial and interstadial deposits is presented on maps, and deposits of this type are also included in some of the selected Quaternary stratigraphical examples. This compilation may be used simply as it is referred to, i.e., an overview of the Quaternary deposits of Norway, but may also be used to consider the overall glacial impact in Norway, with erosion and deposition during the Quaternary (last *c.* 2.6 million years). For example:

The distribution of Quaternary glacial deposits on and beyond the shelf (Figures 2–3) indicates three major depocentres, i.e., the Mid-Norwegian shelf and the areas around the outer parts of the Norwegian Channel and the Bjørnøya Trench, including the adjacent trough-mouth fans. Almost all of the Mid-Norwegian glacial deposits are assumed to derive from the Norwegian mainland, whereas a significant or even major part of the products of glacial erosion stored in the two other depocentres may derive from other areas, e.g., the southern North Sea and the Baltic region in the south and the Barents Sea region in the north.

The volume of the glacial deposits offshore the Mid-Norwegian coast amounts to almost 100,000 km^3 (Dowdeswell et al. 2010), which represents erosion of a size similar to a *c.* 500 m-thick sheet of erosional products covering the overall 200,000 km^2 catchment area in Central Norway and Sweden. A sheet of erosional products of at least 300–400 m thickness covering all onshore areas (*c.* 370,000 km^2) of Norway is needed to illustrate the volume of the glacial deposits offshore all parts of Norway.

The glacial deposits offshore Norway have a volume that indicates sediment supply from erosion of the land with average more than 10 cm, probably *c.* 20 cm kyr^{-1} everywhere in Norway during the entire Quaternary. (The real erosion rate is, however, likely to vary between almost 0 cm to almost 1 m kyr^{-1}, since very old ground surfaces are recorded in some mountain areas, whereas a relief of, e.g., more than 2000 m in the Sognefjord area is thought to have been formed mainly during the Quaternary.)

The loose deposits of onshore areas in Norway are strongly dominated by glacial erosional products and amount to less than 1000 km^3 in volume (i.e., 6 m average sediment thickness in 25% of the area and 1 m in the remaining 75%, which give 2 m average sediment thickness everywhere and a total volume of *c.* 820 km^3).

Almost all of *the Quaternary deposits of onshore areas* in Norway are younger than 300 ka and probably >90% is younger than 115 ka. This gives an average net erosional rate of less than 1 cm kyr^{-1} during the two last glaciations, the Saalian and the Weichselian, and about 2 cm kyr^{-1} during the last glaciation.

The difference in net erosional rates based on the volumes of the offshore (*c.* 20 cm kyr^{-1}) and the onshore (1 cm kyr^{-1}) glacial deposits indicates that >90 % of the glacial erosional products are removed from onshore areas and deposited offshore during the Quaternary. This transport and depositional process must have occurred mainly during large glacier extensions as the glaciers grew to ice sheets reaching the shelves, and it demonstrates clearly the effectiveness of the ice sheets as transport agents for the available erosional products. In general terms, the stratigraphical record with more reworked older deposits and more of the sediments that are preserved derive from younger glaciations support strongly this model, which implies that most of the sediments deposited or overrun by the ice during one glaciation are eroded and redeposited or removed during the subsequent glaciation.

From the record of *the Quaternary glaciations in Fennoscandia* it is found that large ice sheets reached the shelves several times during the last 0.5–1.1 Ma, and clearly less frequently before that. The denudation rate of the land surface, which is estimated to be an order of magnitude higher during ice-sheet intervals than during intervals of moderate to minimum ice extension (Dowdeswell et al. 2010), is therefore likely to have increased significantly towards younger ages during the Quaternary.

The recorded *glacial stratigraphy from onshore sites* in Norway reaches from MIS 1 to MIS 10. In addition to dates, counting from the top of stratigraphical sediment units seems to work well in Norway as a method of stratigraphical correlations, at least for the Weichselian glaciation. However, this is not quite in agreement with the general glacial erosion – accumulation history for the entire Quaternary in Norway as described above.

Acknowledgements

The utilised information is from various sources, but focused on data from NGU where this has been possible (NGU-participation in research). In a few cases, both from land and seabed areas this has not been possible and external sources have been included.
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Appendix A

Principles for stratigraphical correlations used in northern and central Fennoscandia

A general stratigraphical model was established for northern Fennoscandia in the Nordkalott Project (Hirvas et al. 1988) (Figure A1) and modified and extended to central Fennoscandia in the Mid-Norden Project (Bargel et al. 1999). The main basis for correlations is general till stratigrahic data and dates of intercalated deposits from interstadial/ice-free intervals (e.g., Hirvas et al. 1988, Bargel et al. 1999). The age assignments are mainly based on luminescence (TL and OSL) dates of sediments from the last interglacial/glacial cycle, and stratigraphical position with respect to the dated units. The age model (Figure 11, main text) for the glacier fluctuations is based on dates from inland sites, with the implication that a TL or OSL dated ice-free unit indicates regional ice-free conditions and not only a minor ice-marginal retreat from a moderate to major ice extension.

The dates are not corrected for shallow traps, which previously were supposed to underestimate the ages, at least those of 100 ka or more with as much as 35% (Olsen et al. 1996a). The need for such corrections is so far not shown to be valid for sediments in Fennoscandia (Olsen in Lokrantz and Sohlenius 2006). However, a possible general underestimation of *c.* 10% of older ages (>100 ka) should clearly be considered, in agreement with recent results from dating accuracy testing of OSL dates from northern Russia (Murray et al. 2007) and farther south in Europe (e.g., in Germany). The luminescence age intervals of the general age model overlap fairly well with most 'warm' isotope stages (Figure 11, main text), and indicate a much better matching with the MIS boundaries than if a general underestimation of ages is included (stippled curve in Figure 11, main text), but MIS 5a seems to be less well represented by luminescence dates. Alternatively, the luminescence age interval representing MIS 5a is 10 kyr younger (underestimated) than the orbitally tuned MIS 5a age boundaries after Martinson et al. (1987).

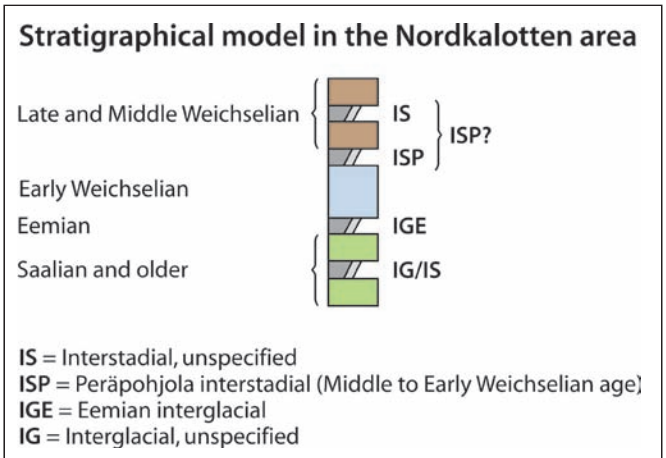


Figure A1: Generalised Quaternary stratigraphical model for northern Fennoscandia (The Nordkalott Project 1986).

The till stratigraphy of Finland, as represented by the till beds named Till 1 (youngest) to Till 6 (oldest), and the intercalated sediment units, is used as a stratigraphical framework for the whole inland area of northern and central Fennoscandia. The updated version of the stratigraphical framework is given in Table A1.

The key unit to recognise for correlation in each studied area is the regional till bed no. 3 (Till 3, which is of Early Weichselian age) counted from the top, with its lithological properties, clast fabric and stratigraphical relationship to the most prominent glacial landforms (for example, this till appears morphologically as NW–SE-oriented megascale drumlins in northern Sweden), and to the other recorded typical regional glacial and nonglacial units. Till 3 is assigned an age of MIS 5d (Hirvas et al. 1988, Olsen 1988, Hirvas 1991, Nenonen 1995, Olsen et al. 1996a) or 5b, as indicated from the studies in the Sokli area, northeastern Finland, by Helmens et al. (2000).

However, based on the complexity of the stratigraphy and variation of the lithology of this region it is tempting to suggest that there are two Till 3 units of which one correlates with 5d and the other with 5b. A two-substage model for Till 3 seems to be the best-fit choice to explain all the reported, relevant observations and may now be regarded as a working hypothesis (Larsen et al. 2005).

Interglacial organic beds (peat, gyttja) have been found between Tills 3 and 4, and they have all been correlated to the last interglacial, the local Tepsankumpu and the regional northwest-European Eemian interglacial (MIS 5e), whereas those found between Tills 4 and 5, at one site only (Naakenavaara, Kittilä), must be older than the Eemian. The interglacial deposits in northern Finland cannot be distinguished merely on a palynological basis (Hirvas 1991), so a correlation based on these beds must also consider their stratigraphical relationships to the regional till beds, and dates (if available).

Organic deposits from two forested Early Weichselian interstadials and one Middle Weichselian interstadial characterised by tundra vegetation are recorded in a stratigraphical succession at Sokli, northern Finland (Helmens et al. 2000). All these deposits are separated and overlain by a till bed. Only inorganic waterlain sediments have been found between Tills 1 and 2 in Finland so far (Hirvas 1991).

The stratigraphical correlation of Tills 1–3 between northern Finland and Finnmarksvidda, northern Norway, is carried out directly based on a collaborative excavation–based fieldwork, using the same methods in both areas (Olsen 1988). Consequently, the correlation between the (youngest) Saalian tills, named Till 4 in northern Finland and Tills 4–5 in northern

Table A1: Stratigraphical framework for northwest Europe, with special emphasis on northern and central Fennoscandia. Modified from Larsen et al. (2005).

		Marine isotope stage	Stratigraphic units/phases used at a national scale		
Stage	**Substage**	**MIS**	*Finland*	*Sweden*	Norway
Weichselian	LGM	2	Tills 1 and 2	Tills 1 and 2	Tills 1 and 2
	Younger Middle Weichelian interstadials	3	Y. MW interst.(?)	Tärendö interst., younger phase (?)	Sargejohka interst., younger phase
	Younger MW stadial	3	Till 2, middle	Till 2, middle (?)	Till 2, middle
	Older MW interstadial	3	MW interstadial, type Sokli	Tärendö interst.	Sargejohka interst., older phase
Average age;	Older MW stadial	4	Till 2, older type Sokli	Till 2, older	Till 2, older
	Odderade interstadial (Early Weichselian)	5a	Y. EW interstadial, type Sokli	Jämtland interstadial, younger phase	Eiravarri interstadial, younger phase
	Younger EW stadial	5b	Till 3, younger, type Sokli	Cold phase	Glacial deformation, type Vuolgamasjohka
	Brørup interstadial	5c	O. EW interstadial, type Sokli	Jämtland interstadial, older phase	Eiravarri interstadial, older phase
	Older MW stadial	5d	Till 3, older (not found at Sokli)	Till 3	Till 3
Eemian interglacial	Eemian interglacial	5e	Tepsankumpu Eemian interglacial	Leväaniemi Eemian interglacial	Eemian interglacial palaeosol, Sargejohka
Saalian *(sensu stricto)*	Warthe and Drenthe stadials	6	Till 4	Till 4	Tills 4 and 5
Saalian Complex *(sensu lato)*	Wacken–Dömnitz–Schöningen interglacial	7	Naakenavaara interglacial	Oje interglacial	Bavtajohka interglacial
	Cold phase; Fuhne	8	Till 5	–	Till 6
	Warm phase; Reinsdorf interglacial	9	Deglaciation phase between tills 5 and 6	–	Palaeosol 5, Sargejohka
	Cold phase (stadial)	10	Till 6	–	Till 7
Holsteinian interglacial	Holsteinian interglacial	11	–	–	Palaeosol 6, Sargejohka
Elsterian	Elsterian	12	–	–	Degl. sediments, unit K, Sargejohka

Norway, seems like a best choice since these tills in both areas occur between interglacial deposits of which the youngest represents the Eemian. The uncertainties are mainly connected to the underlying beds.

Based on counting from the top, stratigraphical relationship to other recorded regional events and luminescence dates the Bavtajohka interglacial from Finnmarksvidda is inferred to correlate with MIS 7, possibly an early part of the stage (MIS 7e) (Appendix B, Table B2) (Olsen et al. 1996a), but a recent re-evaluation of the dates suggests a correlation with a later part, i.e., MIS 7a–7c (Figure 14, main text). Moreover, the stratigraphical position suggests also a correlation with the Naakenavaara interglacial from northern Finland. The overall pollen signature from these interglacials does not disprove such a correlation (Hirvas 1991, Olsen et al. 1996a). If this correlation is correct, then it follows that Till 6 from Finnmarksvidda correlates most likely with Till 5 from northern Finland and, furthermore, that Till 7 from Finnmarksvidda correlates with the Finnish Till 6 (Table A1).

Overview and summary of the Weichselian glacial deposits (described chronologically and from north to south)

Glacial deposits from the Early to Mid Weichselian (MIS 5 to 3)

Tills deposited during a MIS 5d glaciation** (c. **100–110 ka)

No deposits from Weichselian glacial advances prior to LGM 2 have been recorded on Varangerhalvøya, northern and eastern parts, so far, e.g., (1) at Komagelva (Figure 22b, main text), late Saalian–Eemian sand is directly overlain by late Weichselian gravel and LGM 2 or younger till and deglaciation sediments, and (2), at Leirelva, pre-Weichselian gravel is overlain by late Weichselian ice-dammed sediments and LGM 2 or younger till and deglaciation sediments (Olsen et al. 1996a).

MIS 5d ice flowing eastwards and southwards from the mountains deformed Eemian sediments on Finnmarksvidda, but no till from this phase is found so far in this area. The older part of Till 3 in the inland of northern Fennoscandia is assumed to be of MIS 5d age (see above). Maximum ice extension during 5d has probably reached almost, but not completely the inner fjord areas of southern Troms and mid-Nordland, because speleothems grew continuously from the Eemian to *c.* 100 ka in Stordalsgrotta (920 m a.s.l., Stordalen, southern Troms) (Figure 4, main text) and to *c.* 90 ka in mid-Nordland (Fauske–Rana area), and were abruptly ended at these times.

The MIS 5d ice reached almost to Fjøsanger (Bergen) in the west and deposited sediments that demonstrated a glacial climate during the Gulstein stadial (Mangerud et al. 1981b) (Figure 28, main text, silt G), and it deposited till beds (Tills P–Q) at Dalseng (Brumunddalen) in southeast Norway (Figure 32a) (Helle 1978) and probably reached the Oslofjord area.

Tills deposited during a MIS 5b glaciation (*c. *85–90 ka)

The Gardejohka till on Finnmarksvidda, and the younger part of Till 3 in large areas in the inland of northern Fennoscandia, including the Sokli area in northeast Finland, are now reconsidered to have been deposited by a MIS 5b ice (Helmens et al. 2000, Larsen et al. 2005, Olsen 2006) and not a MIS 5d ice as previously suggested (Hirvas et al. 1988, Olsen et al. 1996a). The 5b–ice extension may have reached the innermost fjord areas of mid-Nordland since speleothem growth was interrupted there at *c.* 90 ka. However, continuous speleothem growth from the Eemian to 71–73 ka in caves at 160–220 m a.s.l. in the Fauske–Rana area, indicates that the 5b glacier ended at the fjord heads.

The Bønes till at Fjøsanger, in the coastal area close to Bergen, was deposited by a 5b glacier (Mangerud et al. 1981), which must therefore have reached the sea (Norwegian Channel) in the west. The 5b glacier also deposited tills (Tills K–L) at Dalseng in southeast Norway (Helle 1978) and probably reached the Oslofjord area, similar to the 5d glacier.

Tills deposited during a MIS 4 glaciation (*c. *60–70 ka)

The Vuoddasjavri till on Finnmarksvidda, known as Till 2 in large areas in the inland of northern and central Fennoscandia, was deposited by a MIS 4 ice which extended to the coastal zone in the northwest, as indicated by the interrupted speleothem growths at low altitudes in mid-Nordland.

This ice advance seems to have reached beyond Brumunddalen in southeast Norway, as mentioned before and indicated by a U/Th-dating at 59 ka of the Brumunddalen interstadial peat (age indicates time for closure of previous open circulation and thereby starting the 'U/Th-clock'). A till bed (the Karmøy diamicton) at Bø on Karmøy in southwest Norway is also thought to have been deposited during MIS 4 (Andersen et al. 1983, Sejrup 1987), and this glacier must therefore have reached the sea/Norwegian Channel.

The record of ice-rafted detritus (IRD) off the Norwegian coast shows the highest Early to Mid-Weichselian peak at *c.* 55 ka (Baumann et al. 1995), and the IRD input in sandy mud beyond the shelf break west of the Lofoten islands was intense in the earliest part of the interval 63–54 ka (Laberg and Vorren 2004, and pers. comm. 2007), which suggest that the ice extension during MIS 4–3 reached its maximum off the coast to the west and started to retreat at this time.

Tills deposited during a MIS 3 glaciation, older phase (*c. *44–49 cal ka)

MIS 3 ice extension may have reached the outer coast or even offshore in northern Norway, Mid-Norway, southwest Norway, as well as in southeast Norway where the ice advance may have reached almost to Vendsyssel in Denmark (Houmark-Nielsen and Kjær 2003). In the north, glaciomarine sediments with *c.* 44 cal ka shells from several sites in onshore positions, e.g., at Leirhola (Arnøya), in Løksebotn (Salangen) and at Bleik (Andøya, Møller et al. 1992), indicate considerable glacioisostatic depression in a period with generally low global sea level (e.g., Shackleton 1987). This suggests a considerable ice volume in the northwest at or just before 44 cal ka. Sediments from the deglaciation of this ice are recorded and TL dated to 37–41 cal ka on Finnmarksvidda.

Till beds with shells of age *c.* 44 cal ka are recorded several places in Troms, e.g., at many islands, such as Arnøya (Figures 4 and 23, main text), Vanna, Kvaløya, Grytøya and Hinnøya, and also at other sites in the fjord areas. Similar data exist for Nordland, both from sites on islands and in the fjord districts. For example, shell dates from a complex ice-marginal terrace at Skogreina at the coast close to the Arctic Circle indicate a phase with the ice-margin position located at this site at *c.* 44 cal ka (Olsen 2002). Ice dammed/marine sediments with resedimented shell fragments of age *c.* 50 cal ka may indicate a *c.* 44 cal ka ice advance crossing the Hundkjerka cave in Velfjord, southern Nordland (Figures 4 and 26, main text).

In Trøndelag this ice advance may have reached Vikna in the northwest, as indicated by dates of shells in till (Bergstrøm et al. 1994, Olsen et al. 2001a). In Gauldalen south of Trondheim, Mid-Norway, till beds and intertill sediments, which include sediments with partly decomposed marine fossils, contain also plant remains that have been dated (Figures 4 and 27b, main text). These data indicate ice-margin fluctuations in this area several times around *c.* 44 cal ka, as well as younger fluctuations (Olsen et al. 2001b, c).

Ice-dammed sediments in the cave Skjonghelleren at Møre, farther south along the coast, include a geomagnetic signal (in clay L, Figure 28a, main text) which is correlated with the Laschamp excursion (*c.* 42–47 cal ka). Consequently, the ice advanced beyond the coast at Møre *c.* 44 cal ka. The Sandnes interstadial marine and glaciomarine sediments at high altitudes on Høgjæren in southwest Norway are thought to indicate considerable glacioisostatic depression due to a big ice volume close to this area around 44 cal ka. This ice may have been located onshore, as well as in the Norwegian Channel (Larsen et al. 2000).

At Sørlandet in southernmost Norway, and in southeast Norway subtill sediments with marine fossils/other components at high altitudes include *c.* 38–40 cal ka or older plant remains, which also indicate fast deglaciation of a big *c.* 44 cal ka ice.

The Mesna till from Lillehammer, southeast Norway, has been dated to (framed) between 40 and 44–48 cal ka, which points to a *c.* 44 cal ka age for the associated ice advance, which is thought to be represented by till in a wide area in the inland of southeast Norway (Figure 33b, unit A, main text) and may well have reached the coastal area in the southeast.

Glaciotectonics in the Skara area (northeast of Gothenburg, Sweden) are recorded in reddish-brown clay which is TL and OSL dated to *c.* 43 and 45 cal ka (Ronnert et al. 1992). This means that a glacial event existed in southern Sweden around or after *c.* 43–45 cal ka, and before the southerly ice flows during the Late Weichselian glaciation. This glacial event may therefore correspond to a stadial *c.* 44 cal ka, or a younger event.

Glacial deposits from the Late Weichselian (MIS 2)

Tills from MIS 3/2 and MIS 2 glacial fluctuations (c. 34–17 cal ka)

The sedimentary stratigraphies and glacial fluctuations during this period are thoroughly reviewed by Olsen et al. (2001a, b, c). Therefore, this compilation will only briefly mention data known before the year 2001 (e.g., as those indicated in Figure 16, main text) and mainly concentrate on new information from this period, which includes the biggest Weichselian ice-marginal fluctuations. A new understanding of ice-marginal dynamics is represented by the composite glacial curve after Olsen (1997) (see Olsen et al., 2013, fig. 9bc, which is a synthesis of glaciation curves from nine different transects from inland to shelf). It has as one of its implications that precise dating and age estimates of events are even more important for correlations now than before, a time when most glacial geologists believed in glacial stability, with the ice-margin fluctuations during LGM constrained to the shelf to coast area, i.e., in almost maximum position for thousands of years.

Offshore data, including glacial deposits from more than one phase with the ice margin in the maximum position at the shelf edge during the LGM interval, are very scarce, but still represented both in the north and the south. Off Troms and in the southwest Barents Sea area, Vorren et al. (1990) describe deposits from two maximum phases (LGM I and II), separated by an ice-retreat phase. More precise data, representing more than one LGM advance phase are reported from the Andfjorden area, where both marine and terrestrial sediments indicate at least two major Late Weichselian ice advances, that Vorren and Plassen (2002) describe as Egga I and II (here named LGM 1 and 2) as they infer that both these advances reached the shelf edge. These advances were separated by a complex ice-retreat phase, i.e., an overall ice retreat–ice advance (the Bjerka Substage)–ice retreat phase. The Andfjorden area is located at some distance from the shelf edge and therefore its significance for the glacial maximum on the shelf rely fully on Vorren and Plassen's suggested correlations, which the present authors find reasonable.

At the mouth of the Norwegian Channel in the south (Figure 4, main text), Olsen et al. (2004) speculated from adjacent land data that the ice margin may have reached the maximum position at least three times between 29–18.5 cal ka. This was recently confirmed by Nygård et al. (2007) based on data from the North Sea Fan and a conceptual model where debris flows are initiated by the ice front in the maximum position. They found that this occurred three times during the LGM interval (30–19 cal ka).

Till deposits at the coast, e.g., at Arnøya, northern Norway (Andreassen et al. 1985, and new data recorded by NGU) (Figure 23, main text), and ice-dammed sediments in caves at the coast in west Norway (Larsen et al. 1987, Valen et al. 1996) indicate that a regional ice advance *c.* 33–34 cal ka crossed the coastline. It is supposed to have reached its maximum position at least to around halfway to the shelf edge, similar to that indicated for the glacier during MIS 3 (*c.* 44 cal ka) (Olsen et al. 2001b, c).

At Bø on Karmøy, southwest Norway, the upper till (the Haugesund diamicton, Andersen et al. 1983, Sejrup 1987) has recently been subdivided in at least two different tills with intercalated sediments (Figure 30) (Olsen et al. 2004, Olsen and Bergstrøm 2007). The oldest of these tills is framed in age between 38.5 and 27.9 cal ka, whereas the uppermost part of the upper regional till, which was deposited during a regional phase with W-trending drumlins, is younger than 18.6 cal ka.

Seven additional localities to that from Karmøy and to those with indications of LGM-interval oscillations as reported by Olsen et al. (2001a, b, c) have recently been recorded. These are found in different parts of Norway, i.e., one in the north (Arnelund), two in the central areas (Meråker and Kvikne), one in the southwest (Foss–Eikeland, Figure 31), and three in the southeast (Braarød, Korterød and Børsåsen) (Figure 4, Olsen et al. 2004). The uppermost regional till at all these and previously reported Norwegian LGM oscillation-sites, is generally younger than 19–21 cal ka (Olsen et al. 2001a, b, c, Olsen et al. 2004, Olsen and Bergstrøm 2007, Appendix B, Figure B5).

Tills from MIS 2 and early MIS 1 glacial fluctuations (16 cal ka, and younger)

The last ice retreat from the shelf occurred during the beginning of this period, and the surficial till is of this age in most parts of Norway where the ice sheet produced till during general ice retreat. The oldest post-LGM onshore glacial deposits that are recorded so far are listed in Table B11 (Appendix B) (See also fig. 20 in the accompanying paper of 'Quaternary glaciations', this issue).

Much of the surficial tills proximal to the Nordli–Vuku moraines at the fjord heads in Nordland and Trøndelag (Andersen et al. 1995, Sveian 1997) and the corresponding ice margin in other parts of the country may be of early Holocene (Preboreal) age.

Appendix B

Table B1: Name of all geographical places and sites mentioned in the main text, with geographical positions indicated by numbers (and dots) in Figure B1. (a) Sites arranged successively according to their number, and (b) in alphabethical order.

1	Vardø
2	Leirelva, Komagelva
3	Straumsnesa, Kongsfjorden
4	Varangerfjorden
5	Pasvik
6	Tanafjorden
7	Risvika
8	Finnmark
9	Porsangerfjorden
10	Gamehisjohka
11	Buddasnjarga
12	Balucohka
13	Sargejohka
14	Kautokeino
15	Vuolgamasjohka
16	Vuoddasjavri
17	Iesjavri
18	Alta
19	Lauksundet
20	Leirhola, Arnøya
21	Vanna
22	Troms
23	Lyngsalpene
24	Kvaløya, Slettaelva
25	Arnelund, Balsfjorden
26	Senja
27	Æråsvatnet, Øvre & Nedre Åsvatnet
28	Andøya, Flesen
29	Andfjorden
30	Løksebotn, Salangen
31	Vesterålen
32	Grytøya
33	Hinnøya
34	Lofoten
35	Langøya
36	Narvik
37	Vestfjorden
38	Kjøpsvik, Tysfjorden
39	Steigen
40	Værøy
41	Røst, Trenyken
42	Fauske, Okshola
43	Skogreina

44	Ørnes
45	Meløya
46	Åmøya, Bogneset, Bjærangsfjorden
47	Grønligrotta
48	Hamarnesgrotta
49	Mo i Rana
50	Nordland
51	Røssvatnet
52	Hattfjelldal
53	Fiskelauselva
54	Hundkjerka, Hommelstø, Velfjorden
55	Vikna, Langstrandbakken
56	Namsen
57	Lierne
58	Steinkjer
59	Follafoss
60	Trøndelag
61	Fosenhalvøya
62	Trondheimsfjorden
63	Trondheim
64	Meråker
65	Langsmoen
66	Flora, Drivvollen
67	Byneset, Hangran
68	Grytdal
69	Hauka
70	Skjenaldelva
71	Frøya, Svellingen, Skarsvågen
72	Surna
73	Røros
74	Kvikne, Kvikneskogen
75	Trollheimen
76	Oppdal
77	Møre, Nordmøre, Sunnmøre
78	Ålesund
79	Skjonghelleren
80	Hamnsundhelleren
81	Skorgenes
82	Lesjaskog
83	Galdhøpiggen, Jotunheimen
84	Dovre
85	Rondane
86	Jutulhogget

87	Glåmdalen
88	Venabu
89	Vinstra, Sorperoa, Gudbrandsdalen
90	Nordfjord
91	Sognefjorden
92	Kollsete
93	Kroken, (Inner Sogn)
94	Møklebysjøen
95	Lillehammer, Mesna, Stampesletta
96	Øvre Åstbrua
97	Dokka
98	Hunndalen
99	Dalseng, Brumunddal
100	Rokosjøen, Rokoberget
101	Bergen, Herdla, Fjøsanger
102	Halsnøy
103	Hardangerfjorden
104	Haugesundshalvøya
105	Karmøy, Bø
106	Rogaland
107	Jæren, Lågjæren, Høgjæren
108	Foss-Eikeland
109	Lysefjorden
110	Hardangervidda
111	Mårvatn, Mår
112	Møsvatn, Førnes, Hovden
113	Gardermoen (airport)
114	Oslo
115	Ås, Ski
116	Passebekk
117	Herlandsdalen, Rundhaugen
118	Oslofjorden
119	Halden
120	Hvaler
121	Jomfruland
122	Skagerak
123	Sørlandet
124	Lista

Table B1: (b) in alphabethical order.

Table B2: Correlation chart for the late Middle Pleistocene stratigraphy from Norway, Germany, and France, and the marine oxygen isotope stratigraphy (MIS). Modified from Olsen et al. (1996a) and Larsen et al. (2005).

Northern Norway Sargejohka (Olsen et al. 1996, Olsen 1998)	**Northern Germany** Wacken (Menke 1968, a.o.)	**Southern Germany** Schöningen (Urban 1997, Thieme 1999)	**Southern France** Velay (Reille et al. 1998)	**MIS**
Weichselian	Weichselian	Weichselian	(Last glaciation, local ice cap)	2–5d
EEMIAN PALEOSOL	EEMIAN	EEMIAN	RIBAINS INTERGLACIAL	5e
Stadial (T4) -	Saale (Drenthe)	Saale (Drenthe)	Costaros glacial	6
interstadial -		BUDDENSTEDT 2	BOUCHET III	7a
stadial (T5)			Belvezet stadial	7b
sediment complex		BUDDENSTEDT 1	BOUCHET II	7c
			Bonnefond stadial	7d
BAVTAJOHKA	WACKEN	SCHÖNINGEN	BOUCHET I INTERGLACIAL	7e
Glacial (T6)	Fuhne (?)	Fuhne	Charbonnier glacial	8
interstadial			AMARGIER; with ash bed	9a
sediment			Monteil stadial	9b
complex			USSEL	9c
			Cayres stadial	9d
PALEOSOL 5		REINSDORF	BOUCHET I INTERGLACIAL	9e
Glacial (T7)			Bargette glacial	10
		MISSAUE 2 + SU A	JAGONAS 2	11a
			Pradelle 2 stadial	11b
		MISSAUE 1	JAGONAS 1	11c
			Pradelle 1 stadial	11d
PALEOSOL 6	HOLSTEINIAN	HOLSTEINIAN	PRACLAUX INTERGLACIAL	11e
Elsterian deglaciation				
sediments; Sargejohka unit K				
(no till recorded)	Elsterian	Elsterian	Barges glacial	12

Table B3: Maxima in glacier extension of the Fennoscandian ice sheet during the Middle to Late Pleistocene. Late Pleistocene included for comparison. After Larsen et al. (2005).

Glacial Stage	Substage	MIS (marine isotope stage), age (in calendar ka)	Maximum extension–Ice marginal position			
			North	West	South	East
Weichselian	1. YD (Younger Dryas)	Late MIS 2, 11.5–13	Coast, Norway	Coast, Norway	Coast, Norway, and south Sweden inland	South Finland–northwest Russia
	2. LGM 2 (Last glacial max. 2)	MIS 2, 19 (and 18?)	Outer shelf area	Shelf edge–the Norwegian channel	Central Denmark–north Germany–central Poland	Russia, close to Arkhangelsk
	3. LGM 1 (Last glacial max. 1)	MIS 2/3, 26	Shelf edge	Shelf edge–the Norwegian channel	Central Denmark–north Germany–central Poland	NW Russia (?)
	4. Late M.-Weichselian	MIS 3, 32	Inner shelf area	Shelf, possibly shelf edge	Southern Sweden	Finland–NW Russia (?)
	5. Middle Weichselian	Middle MIS 3, 45	Inner shelf area	Middle shelf area	South Sweden–east Denmark–north Germany–Latvia–Estland	NW Russia (?)
	6. Early M.-Weichselian	MIS 4, 65	Outer coastal zone	Outer coastal zone, possibly shelf and Norw. Channel	South Sweden–east Denmark–north Germany–Latvia–Estland	NW Russia (?)
	7. Late E.-Weichselian	MIS 5b, 85	Inner fjord area	Outer coastal zone	Southern Sweden	Pudasjarvi moraines, north Finland–NW Russia
	8. Early E.-Weichselian	MIS 5d, 110	Inner fjord area	Outer fjord area	Gulf of Bothnia	Lapland moraines, north Finland
Saalian (sensu stricto)	9. Younger Late Saalian *(sensu lato)*	MIS 6, 140	Shelf edge	Shelf edge–the Norwegian Channel	South Germany–south Poland–Bela Russia	NW Russia, east of Arkhangelsk
	10. Older Late Saalian *(sensu lato)*	MIS 6, 185	Coast, possibly shelf	Shelf edge–the Norwegian Channel	South Germany–south Poland–Bela Russia	NW Russia, east of Arkhangelsk
Saalian Complex	11. Younger Middle Saalian *(s.l.)*	MIS 8, 260	Shelf, possibly shelf edge	Shelf (?)–the Norw. Ch. (?)	Southern Baltic basin area (?)	Finland–NW Russia (?)
	12. Older Middle Saalian *(s.l.)*	MIS 8(?), 280–320(?)	Coast, possibly shelf area	Middle shelf area	Southern Baltic basin area (?)	Finland–
	13. Early Saalian *(sensu lato)*	MIS 10(?), 350(?)	Shelf, possibly Shelf edge	Shelf edge–the Norw. Channel	Southern Baltic basin area (?)	Finland–NW Russia (?)
(Elsterian?)	14. Pre-Saalian	MIS 12(?), 430(?)	Coast, possibly shelf area	Shelf edge– the Norwegian Channel	South Germany–south Poland–Bela Russia	NW Russia, east of Arkhangelsk

Table B4: N–S distribution of stratigraphic sites.

Number of localities from various age intervals, in (A) North Norway, (B) Mid-Norway (between Mosjøen and Ålesund), and (C) South Norway

Age	A	B	C	Total
Pre-Eemian	6	2	3	11
Eemian	11	5	11	27
Early Middle Weichselian/ Early Weichselian	12	2	9	23
Early Middle Weichselian	15	8	10	33
Late Middle Weichselian	14	17	18	49
LGM interval	11	17	13	41
Post LGM– pre Older Dryas	4	6	2	12
Lateglacial	common	common	common	many
Holocene	common	common	common	many

Table B5: Pre-Eemian sites, most of these (8 of 11 sites) include Eemian sediment or palaeosol as an upper reference horizon.

Locality		Stratigraphy	Comments
1.	Leirelva, Finnmark, Northern Norway	Late Weichsel. till and sediment/ older sediments	TL-dating: 370 ka => possibly Saalian age sediment
2.	Gamehisjohka, Finnmarksvidda	Eemian sediment/till/sand	Till and sand of probably Saalian age (MIS 6)
3.	Buddasnjarga, Finnmarksvidda	Eemian sediment/till	Till of probably Saalian age
4.	Sargejohka, Finnmarksvidda	Eemian palaeosol/multiple tills and sediment units	Tills and sediment of pre-Eemian, probably MIS 6–11 age
5.	Vuoddasjavri, Finnmarksvidda	Eemian sediment/multiple tills and sediment units	Tills and sediments of pre-Eemian, probably Saalian age
6.	Vuolgamasjohka, Finnmarksvidda	Eemian palaeosol/sediment/ till	Till of pre-Eemian, probably Saalian age (MIS 6)
7.	Skarsvågen, Frøya, Mid-Norway	Eemian marine sediment/ till	Till of pre-Eemian, probably Saalian age (MIS 6)
8.	Ålesund (e.g., Godøya and Vigra)	Multiple tills and sediment units overlain by Early Weichsel. sed.	Tills and sed. of pre-Early Weichsel, probably of pre-Eemian age
9.	Fjøsanger, Western Norway	Eemian marine sediment/till (Paradis till)	Till of pre-Eemian, probably Saalian age (MIS 6)
10.	Grødeland, Jæren, Southwest Norway	Multiple tills and sediment units with marine fossils	Tills and sed. of MIS 6–9 age, based on AAR analyses
11.	Mesna, Lillehammer, Southeast Norway	Pre-Holocene interglacial palaeosol/till (Skjellerud till)	Till of pre-Eemian, probably Saalian age (MIS 6)

Localities:	Geographical area:	References:
1, 2, 3, 4, 5, 6	Northern Norway	Olsen et al. (1996a)
7	Mid-Norway	Aarseth (1995)
8	Mid-Norway	Landvik and Mangerud (1985), Jungner et al. (1989)
9	Southern Norway	Mangerud et al. (1981b)
10	Southern Norway	Janocko et al. (1998)
11	Southern Norway	Olsen (1985b, 1998)

Table B6: All reported Eemian sites from onland positions in Norway, and selected offshore Eemian sites.

Locality	Dating material	Methods	Comments	References:
1. Komagelva	sand, subtill position	OSL	117 and 123 ka => Eemian	Olsen et al. (1996a)
2. Vuolgamasjohka	palaesol in sand, subtill position	TL	stratigraphy and palaeosol indicate Eemian age	Olsen et al. (1996a)
3. Vuoddasjavri	sand, subtill position	TL	114, 123, 126 and 134 ka => Eemian age	Olsen et al. (1996a)
4. Sargejohka	palaeosol (poszol) in sand, subtill position	TL	>50 ka, interglacial pollen => probably Eemian	Olsen et al. (1996a)
5. Gamehisjohka	gyttja, sand, subtill position	C14, pollen	Eemian sediment/multiple tills and sediment units	Olsen et al. (1996a)
6. Buddasnjarga	palaeosol in till, subtill position		correlation to locality 4 and 5 => probably Eemian	Olsen et al. (1996a)
7. Slettaelva	foraminifers in till, resedimented	AAR	probably Eemian, from AAR–age	Vorren et al. (1981)
8. Stordalsgrotta, cave	speleothem	U/Th	Eemian, from U/Th–age	Lauritzen (1995)
9. Okshola, cave	speleothem	U/Th	Eemian, from U/Th–age	Lauritzen (1995)
10. Gammalmunnåga	shells in till, resedimented	C14, AAR	>48 ka (C14) and AAR–result indicate Eemian age	Olsen et al. (2001a), Olsen (2002)
11. Hamarnesgrotta and Grønligrotta, caves	speleothems	U/Th	Eemian, from U/Th–age	Linge et al. (2001)
12. Skarsvågen	pollen and shells in subtill sediments	AAR	stratigraphy and AAR–age => probably Eemian age	Aarseth (1990, 1995)
13. Follafoss	shells in till, resedimented, and in intratill sediment	C14	>48 ka and high elevation => probably early Eemian	Olsen et al. (2001a)
14. Drivvollen and Langmoen	foraminifers in silt, subtill position		interglacial foraminifers => probably Eemian age	Olsen et al. (2002)
15. "Møre", several sites	shells in till, resedimented	AAR	Eemian age, from AAR analyses	Mangerud et al. (1981a)
16. Kroken, Sogn	pollen in till/ subtill sediments	pollen (picea)	pollen and stratigraphic pos. => Eemian age	Vorren (1972)
17. Fjøsanger	pollen and shells in subtill marine sed.	AAR, TL	warm interglacial flora and fauna => Eemian age	Mangerud et al. (1981b)
18. Vinjedalen, Vossestrand	peat (2 cm thick), subtill position	pollen (picea)	stratigraphy and pollen => Eemian age	Sindre (1979)
19. Bø, Karmøy	shells in marine sediments, subtill position	AAR	warm fauna and AAR–age => Eemian age	Andersen et al. (1983), Sejrup (1987)
20. SW Norway, Haugesunds-halvøya and Jæren	pollen and shells in till/ subtill sed.	pollen, AAR	interglacial flora/ fauna => probably Eemian age	Vorren and Mangerud (2006)
21. Hovden	pollen in till/ subtill sediments	pollen (picea)	warm interglacial flora and stratigraphic pos. => Eemian	Vorren and Roaldset (1977)
22. Lågen–Gausa deltaet	shell (glycimeris), resedimented in delta	C14	>50 ka and warm interglacial shell => Eemian age	Olsen and Grøsfjeld (1999)
23.			stratigraphy and character of palaeosol => Eemian age	Olsen (1985b, 1998)
24. Central SE Norway, several sites	sand, subtill position, resedimented	TL, OSL	110–140 ka and strat. pos. => probably Eemian sed.	Myklebust (1991), (O.F. Bergersen, unpubl. 1991)
Offshore sites:				
102. Smørbukk, Mid-Norwegian shell	Strata including Eemian beds			Sættem et al. (1993)
2501 Statfjord, North Sea Plateau	Strata including Eemian beds			Feyling-Hanssen (1981)
5.1/5.2 Troll, Norwegian Channel	Strata including Eemian beds			Sejrup et al. (1991, 1994)
Eem-site in Denmark:				
Apholm	Strata including Eemian beds			Knudsen and Sejrup (1999)

Table B7: Early Weichselian and early Middle Weichselian sites (with sediments and tills). Average number of younger (overlying) stadials/glacial units is c. 2.65 (n=23).

Locality		Stratigraphy	Comments	Overlying glacial units	References:
1.	Gamehisjohka, Finnmarksvidda	Till overlying Eemian palaeosol and overlain by LW* till and deglaciation sediments	=> till of probably EW* to early MW* age	1	Olsen et al. (1996a)
2.	Buddasnjarga, Finnmarksvidda	Till overlying Eemian palaeosol and overlain by LW till and deglaciation sediments	=> till of probably EW to early MW age	2	Olsen et al. (1996a), (L.Olsen, unpubl. 1997)
3.	Sargejohka, Finnmarksvidda	Till overlying Eemian palaeosol and overlain by MW–LW sediments	=> till of EW to early MW age	3	Olsen et al. (1996a)
4.	Vuoddasjavri, Finnmarksvidda	Till overlying TL-dated Eemian beds and overlain by MW–LW tills	=> till of probably EW to early MW age	4	Olsen et al. (1996a)
5.	Vuolgamasjohka, Finnmarksvidda	Till overlying TL-dated Eemian sediment and overlain by MW–LW tills/sediments	=> till of EW age	4	Olsen et al. (1996a), Lyså and Corner (1993)
6.	Kautokeino, Finnmarksvidda	Palaeosol with ice wedge cast between two TL-dated pre-MW and pre-LW degl. units	=> degl./glacial sediments of EW to early MW age	3	Olsen et al. (1996a), Olsen and Often (1996)
7.	Vuoskujavri, Finnmarksvidda	Till overlying sand with interglacial pollen and overlain by MW–LW tills	=> till of probably EW to early MW age	2	Olsen et al. (1996a), Olsen and Often (1996)
8.	Ås'kal, Finnmarksvidda	Deglaciation sediments with interglacial pollen overlain by MW and LW tills	=> probably degl./glacial sediments of EW age	2	Olsen and Hamborg (1984), (L.Olsen, unpubl.)
9.	Slettaelva, Kvaløya, Troms	Till overlain by MW–LW sediments and tills, with marine shells	stratigraphy and AAR of shells => EW to MW age	3	Vorren et al. (1981)
10.	Storelva, Grytøya, southern Troms	Till overlain by MW intratill sediments with C14- and AAR dated shells	stratigraphy and AAR of shells => EW to MW age	2	Olsen and Grøsfjeld (1999), Olsen et al. (2001a)
11.	Skogreina, mid Nordland	Ice marginal complex from MW and LW, reworked till/sediment, AAR dated shells	stratigraphy and AAR of shells => EW till/sediment	2	Olsen et al. (2001a)
12.	Gammalmunnåga, mid Nordland	Reworked till/sediment overlain by LW till, C14- and AAR dated shells	stratigraphy and AAR of shells => EW	1	Olsen et al. (2001a)
13.	Skarsvågen, Frøya, Trøndelag	Till overlying in situ Eemian marine and terrestrial sediments, AAR dated shells	stratigraphy and AAR of shells => probably EW–MW	1	Aarseth (1990, 1995)
14.	Folldal, northern Hedmark	Till overlain by MW–LW tills/sediments, C14-and U/Th dates from overlying sediment	stratigraphy and dates from overlying units => EW–MW till	3	Olsen et al. (2001), (L.Olsen, unpubl.)
15.	Skjonghelleren, Valderøy, Møre	Ice-dammed sediments underlying sed. with 50–60 ka (U/Th) old speleothems	stratigraphy and U/Th age of overlying units => EW sed.	4	Larsen et al. (1987)
16.	Dalseng, southern Hedmark	Tills underlying Brumunddalen interstadial peat of EW–MW age	Stratigraphy => EM tills	3	Helle et al. (1981)
17.	Fjøsanger, western Norway	Bønes till, overlying in situ Eemian and EW sediments, AAR dated shells from the till	Stratigraphy and dates => probably EW–MW age	2	Mangerud et al. (1981)
18.	Bø, Karmøy, southwestern Norway	Karmøy diamicton (till) between Bø and Torvastad interstadials	Stratigraphy => EW till	3	Andersen et al. (1983), Sejrup (1987)
19.	Auestad, Jæren, southwestern Norway	Skretting diamicton (till) of EW age, correlated with Auestad diamicton (till) 4	Stratigraphy and regional correlation => EW age of till	2	Janocko et al. (1998), Larsen et al. (2000)
20.	Skretting, Jæren, southwestern Norway	Skretting diamicton (till), EW age, overlying LS* Auestad clay, AAR dated shells	Stratigraphy and dates => EW age of till	2	Stalsberg et al. (1999), Larsen et al. (2000)
21.	Mår, Hardangervidda	Mår lower till, correlated with Hovden lower till and overlain by OSL dated MW–LW sediments	Stratigraphy and dates => EW–MW age of till	4	Vorren (1979), Haug (2005)
22.	Hovden, Hardangervidda	Hovden lower till, between OSL dated MW–LW sediments and Hovden clay of Eemian age	Stratigraphy and dates => EW–MW age of till	4	Vorren (1979), Haug (2005)
23.	Førnes, Hardangervidda	Førnes lower till, correlated with Hovden lower till	Stratigraphy and regional correlation => EW age of till	4	Vorren (1979), Haug (2005)

*LW = Late Weichselian; MW = Middle Weichselian; EW = Early Weichselian; and LS = Late Saalian

*Table B8: Early Middle Weichselian sites (with sediments and tills). Average number of younger (overlying) stadials/glacial units is c. 1.72 (n=33). *Mainly ^{14}C-dates, ages are in ^{14}C yr BP (± 1std). Other dates are also included (U/Th, AAR) and these are in cal yr BP.*

Locality					
1, Vanna, Troms	Till with reworked till/sed. with shells	39,495 +/-870	Probably resedimented early MW till/sediments	1	Olsen (2004), (L.Olsen, unpubl.)
2, Leirhola I-II, Arnøya, Troms	Till with reworked till/sed. with shells	48,635 +2595/-1960 44,755 +1745/-1435	Probably resedimented early MW till/sediments	2	Olsen (2004), (L.Olsen, unpubl.)
3, Kvalsundet, Kvaløya, Troms	Till with reworked sediments/ shells	40,600 +2100/-1700	Possibly MW till	1	Vorren (1979)
Kvaløya, Troms	sediments/shells		and sediments		
4, Slettaelva, Kvaløya, Troms	Till with reworked sediments/ shells	41,900 +2800/-2200	Possibly MW till and sediments	2	Vorren et al. (1981)
5, Lavangsdalen and Arnelund	Till with reworked sediments/ shells	c. 44,000 > 48,300 (+/- 2std)	Possibly MW till and sediments	2	Olsen (2004), (L.Olsen, unpubl.)
6, Bleik, Andøya, Nordland	Glaciomarine sed. with reworked sed.	> 40,000	Ice proximal milieu, possibly MW age	0	Møller et al. (1992)
7, Løksebotn, southern Troms	Tills/intratills sed., with reworked sed.	44,560 +/-2000 47,815 +2305/-1790	Possibly MW tills and sediments	2	Olsen (2004), (L.Olsen, unpubl.)
8, Storelva, Grytøya, Troms	Glaciomarine sed. with reworked sed.	41,660 +/-1500	Ice proximal milieu, possibly MW age	1	Olsen and Grøsfjeld (1999)
9, Mågelva II, Hinnøya	Till with reworked sediments/ shells	45,560 +/-2400	Possibly MW till and sediments	2	Olsen and Grøsfjeld (1999)
10, Raudskjer, Hinnøya	Till with reworked sediments/ shells	42,260 +1165/-1020	Possibly MW till and sediments	1	Olsen (2004), (L.Olsen,
11, Kjøpsvik, cave Nordland	Bone underlying ice-dammed sed.	41,120 +1480/-1250	Ice proximal milieu, possibly MW age	2	Lauritzen and Nese (1996)
12, Grytåga, Nordland	Till with reworked sediments/ shells	41,460 +/-900	Possibly MW till and sediments	1	Olsen et al. (2001a)
13, Gammalmunnåga, Nordland	Till with reworked sediments/ shells	> 44,800	Possibly MW till and sediments	1	Olsen (2002), (L.Olsen, unpubl.)
14, Bogneset, Nordland	Tills/intratills sed., with reworked sed.	40,025 +/-965	Possibly MW tills and sediments	3	Olsen (2002), (L.Olsen, unpubl.)
15, Rana, cave Nordland	Concretions in sediments	46,560 +2700/-2000	Possibly MW degl. sed. and interstadial	0	Olsen et al. (2001a)
16, Hundkjerka, cave Nordland	Glacial diamiction, reworked sed./ shells	46,340 +/-1620	Possibly MW interstadial and ice advance	2	(L.Olsen, unpubl.)
17, Vikna, Trøndelag	Till with reworked sediments/ shells	>40,000	Possibly MW till and/ or MW reworked sed.	1	Bergstrøm et al. (1994)
18, Nordli, Lierne Trøndelag	Intratill sediments with plant fragments	41,000 +3000/-2000	Possibly MW till and/ or MW reworked sed.	2	Olsen et al. (2001a), (L.Olsen, unpubl.)
19, Follafoss I-II Trøndelag	Intratill sediments with shell fragments	46,905 +/-4020 47,565 +/-4680	Possibly MW till and/ or MW reworked sed.	1	Olsen et al. (2001a)
20, Svellingen, Frøya Trøndelag	Till with reworked sediments/ shells	42,400 +1280/-1110	Possibly MW till and/ or MW reworked sed.	1	Aarseth (1990)
21, Grytdal, Trøndelag	Tills/intratills sed., with plant remains	41,800 +1000/-1100	Possibly MW tills and/ or MW reworked sed.	3	Olsen et al. (2001a)
22, Ertvågøya, Møre	Till with reworked sediments/ shells	41,300 +3130/-2240	Possibly MW till and/ or MW reworked sed.	1	Follestad (1992)
23, Skjonghelleren, cave, Møre	Speleothem overlain by ice-dammed sed.	55,700 +/-4000 (U/Th)	MW ice proximal env. and MW interstadial	3	Larsen et al. (1987)
24, Gråbekken, Folldal, Hedmark	Concretions in sub till sediments	41,300 +900/-1000	Possibly MW ice proximal env./ice-free cond.	2	Olsen et al. (2001a), (L.Olsen, unpubl.)
25, Kollsete, Sogn and Fjordane	Gyttja overlain by ice-dammed subtill sed.	43,800 +3700/-2500	MW interstadial and ice proximal conditions	1	Aa and Sønstegaard (1997)
26, Fåvang, Gudbr.d. central south Norway	Mammoth bone from subtill sed., sandur	45,300 +/-2900 (U/Th)	Possibly MW ice proximal env./ice-free cond.	2	Idland (1992)
27, Sæter, Søre Ål, Lillehammer	Mammoth bone from till/ reworked sediments	45,400 +1500/-1200 42,400 +/-500 (U/Th) 52,300 +/-900 (U/Th) 53,900 +/-900 (U/Th)	MW interstadial sed. and MW till	3	Heintz (1974) Idland (1992) Idland (1992) Idland (1992)
28, Øvre Åstbrua, central south Norway	Subtill sediments with plant remains	>48,000 and >50,000	Stratigraphy, dates and pollen => MW or older	2	Haldorsen et al. (1992)
29, Rokoberget, southeast Norway	Subtill sediments with plant remains	47,000 +4000/-3000	Stratigraphy, dates and pollen => MW	2	Rokoengen et al. (1993), Olsen and Grøsfjeld (1999)
30, Hardangervidda, southern Norway	Intratill sediments	c. 40,000 (OSL)	Possibly MW till and/ or MW ice-proximal sed.	3	Haug (2005)
31, Bø, Karmøy, soutwest Norway	Bø interstadial sed. overlain by sed./tills	c. 50,000–55,000, from C14 and AAR data	Early MW interstadial and MW glacial sed.	3	Andersen et al. (1983), Sejrup (1987)
32, Oppstad, Jæren soutwest Norway	Subtill sediments with shells/ forams.	41,300 +6200/-3500 AAR indicates MW age	Possibly MW till and/ or MW interstadial sed.	2	Andersen et al. (1991), Andersen et al. (1987)
33, Høgemork, Jæren southwest Norway	Subtill sediments with shells/ forams.	AAR indicates MW age	Possibly MW till and/ or MW interstadial sed.	3	Andersen et al. (1987)

*) See Table caption

*Table B9: Late Middle Weichselian sites (with sediments and tills). Average number of younger (overlying) stadials/glacial units is c. 1.39 (n=49). [1]Ages are shown in ^{14}C yr BP (± 1std) for ^{14}C dates and in cal yr BP for other dates. *BPR= Bulk plant remains, **Sargejohka interstadial, ***Ørnes interstadial (coast) = Hattfjelldal interstadial II (inland), ****Stratigraphy including known palaeomagnetic signals, *****Trace amounts of marine organisms/fossils (e.g., dinocysts).*

Locality	Dating[1]	Lab. Number	Material/ comments	Overlying glacial units	References
1, Skjellbekken, Finnmark	C14: 34,000 ±600	UtC 4039	BPR*, in subtill sediments	2	Olsen et al. (2001a)
2, Sargejohka, Finnmark	C14: 35,100 ±1600	Ua 319	BPR, in subtill sediments (Sargejohka interstadial**)	1	Olsen et al. (1996a)
3, Kautokeino, Finnmark	TL: 37,000 ±5000 TL: 41,000 ±5000	R-823820a R-823820b	Sand, resedimented interstadial** material	0	Olsen (1988), Olsen et al. (1996a)
4, Arnøya, Troms	C14: 27,400 ±1500 C14: 29,000 ±4200	T-3509 T-4020	Shell, in till and subtill sediments	2	Andreassen et al. (1985)
5, Bøstranda, Langøya, Nordland	C14: 39,150 ±900	T-3942	Shell, in till	1	Rasmussen (1984)
6, Kjøpsvik, cave Nordland	C14: 31,160 ±300 U/Th: 36,000 ±3600	TUa-489 ULB-863	Bone, in sediments under lying ice-dammed sed.	1	Nese (1996), Nese and Lauritzen (1996)
7, Urdalen, Nordland	C14: 27,580 ±220	UtC 8459	BPR, in subtill sediments	2	Olsen et al. (2001a)
8, Hakvåg, Nordland	C14: 36,200 ±500 C14: 39,200 ±700	UtC 13556 UtC 13555	Shell, in till	1	(L.Olsen, unpubl.)
9, Grytåga, Nordland	C14: 35,400 ±500	UtC 5557	BPR, reworked sediments in till	1	Olsen et al. (2001a)
10, Risvasselva, Nordland	C14: 36,800 ±600	UtC 5558	BPR, in subtill sediments	2	Olsen et al. (2001a)
11, Åsmoen, Nordland	C14: 28,355 ±235 C14: 29,075 ±370	TUa-567 TUa-1094	Shell, in subtill sediments (Ørnes interstadial***)	1	Olsen (2002)
12, Bogneset, Nordland	C14: 28,355 ±430 C14: 35,940 ±1455	TUa-1240 TUa-1239	Shell, in subtill sediments***	2	Olsen (2002)
13, Kjeldal, Nordland	C14: 33,700 ±400 C14: 35,800 ±600	UtC 8312 UtC 8311	Shell, in subtill sediments	2	Olsen (2002)
14, Luktvatnet, Nordland	C14: 30,600 ±300	UtC 4715	BPR, reworked sediments in till	1	Olsen et al. (2001a)
15, Fiskelauselva, Nordland	C14: 28,000 ±500 C14: 29,400 ±500	UtC 2215 UtC 3466	BPR, in subtill sediments (Hattfjelldal interst. I, II)	2	Olsen (1997), Olsen et al. (2001a)
16, Røssvatnet, Nordland	C14: 29,700 ±500 C14: 31,000 ±500	UtC 3469 UtC 3468	BPR, in subtill sediments (Hattfjelldal interst. I, II)	1	Olsen et al. (2001a)
17, Hattfjelldal, Nordland	C14: 25,370 ±170 C14: 34,900 ±400	UtC 4721 UtC 4722	BPR, in subtill sediments (Hattfjelldal interst. I, II)	2	Olsen et al. (2001a)
18, Langstrandbakken, Nord-Trøndelag	C14: 36,950 ±2700	T-12564	Shell, in till	2	Olsen et al. (2001a)
19, Namskogan, Nord-Trøndelag	C14: 28,700 ±400	UtC 3465	BPR, reworked sediments in till	1	Olsen et al. (2001a)
20, Gartland, Nord-Trøndelag	C14: 28,000 ±200	UtC 4719	BPR, reworked sediments in till	1	Olsen et al. (2001a)
21, Sitter, Nord-Trøndelag	C14: 30,200 ±400	UtC 2103	BPR, reworked sediments in till	2	Olsen et al. (2001a)
22, Sæterelva, Sør-Trøndelag	C14: 39,140 ±900	TUa-1238	Shell, in till	1	Olsen et al. (2001a), Olsen and Riiber (2006)
23, Nesavatnet, Sør-Trøndelag	C14: 36,815 ±590	TUa-2526	Shell, in till	1	(L.Olsen, unpubl.)
24, Skjenaldelva, Sør-Trøndelag	C14: 33,620 ±470 C14: 34,155 ±620	TUa-2996 TUa-2997	Shells (forams.) in subtill sediments	2	(L.Olsen, unpubl.)
25, Renåa, Sør-Trøndelag	C14: 28,700 ±300 C14: 31,600 ±400	UtC 5549 UtC 5552	BPR, reworked sediments in till	2	Olsen et al. (2001a)
26, Stærneset, Sør-Trøndelag	C14: 25,240 ±180	UtC 5556	BPR, reworked sediments in till	2	Olsen et al. (2001a)
27, Grytdal, Sør-Trøndelag	C14: 28,400 ±300 C14: 37,200 ±600	UtC 5564 UtC 5560	BPR, reworked sediments in till	2	Olsen et al. (2001a)
28, "Ålesund interstadial" sites, Møre	C14: 35,700 ±1100 (Eidsvik)	T-2657	Shells, in till, shells of various ages, also late MW	1	Mangerud et al. (1981a)
29, Kortgarden, Møre	C14: 26,940 ±670	T-7281	Shell, in till	1	Follestad (1990)
30, Hamnsundhelleren, cave, Møre	C14: 24,387–31,905	Also palaeomag. signals	Bones in sediments**** underlying ice-dammed sed.	1	Valen et al. (1996) (several dates)
31, Skjonghelleren, cave, Møre	C14: 28,900–34,400 U/Th:23,900–28,000	Also palaeomag. signals	Bones in sediments underlying ice-dammed sed.	1	Larsen et al. (1987) (several dates)

Locality	Dating[1]	Lab. Number	Material/ comments	Overlying glacial units	References
32, Gråbekken, Folldal Hedmark	C14: 32,520 ±650 C14: 36,300 ±600	T-3556B UtC 4724	BPR and concretions in subtill sediments	1	Thoresen and Bergersen (1983), Olsen et al. (2001a)
33, Haugalia, Oppland	TL: 42,000 ±4000 (C14, U/Th, AAR)	R-897010, also OSL	Mammoth bones in subtill sediments	1	Bergersen and Garnes (1981), Idland (1992), (L.Olsen, unpubl.)
34, Sorperoa, Oppland	TL: 37,400–40000 (range of 4 dates)	R-…, also OSL	Aeolian sand, age supposed older than youngest till	0	Bergersen et al. (1991), (S.O.Dahl, pers. comm. 2004)
35, Fåvang, Oppland	TL: 32,000 ±3000 U/Th: 45,300 ±2900	R-…, also OSL	Mammoth bones in subtill sediments	1	(O.F.Bergersen, pers. comm. 1991), Idland (1992), (L.Olsen, unpubl.)
36, Stampesletta, Oppland	C14: 32,300 ±500	UtC 1965	BPR, in reworked subtill sediments	2	Olsen et al. (2001a)
37, "Øst-Jotunheimen", Oppland	C14: 26,000 ± …	…	BPR in subtill sediments	1	(S. Sandvold, pers. comm. 1997)
38, Dokka, Oppland	C14: 26,800 ±400	UtC 3462	BPR in subtill sediments	2	Olsen et al. (2001a), (L.Olsen, unpubl.)
39, Mesna, Lillehammer Oppland	C14: 31,500 ±700 C14: 36,100 ±900	UtC 2217 UtC 1964	BPR in subtill sediments	2	Olsen (1985b, 1998)
40, Hunndalen, Gjøvik, Oppland	C14: 34,200 ±600	UtC 14607	BPR in subtill sediments	1	(L.Olsen, unpubl.)
41, Rokoberget, Oppland	C14: 33,800 ±800 C14: 47,000 ±4000	UtC 1963 UtC 1962	BPR in subtill sediments, traces of marine fossils*****	1	Rokoengen et al. (1993a), Olsen and Grøsfjeld (1999)
42, Møsvatn sites, Hardangervidda	OSL:35,000 ±3000 OSL:41,000 ±4000	R-33507 R-33508	Subtill sand	1	Vorren (1979), Haug (2005)
43, Hovden & Grytkilvika, Hardangervidda	OSL:32,000, 35,000 and 48,000 ±4000	R-33502,33503 and 33510	Subtill sand	2	Vorren (1979), Haug (2005)
44, Bø, Karmøy, Rogaland	C14: 26,510 ±240 C14: 34,000 ±500	TUa-5267 UtC 13562	Shell, in till	2	Ringen (1964), Andersen et al. (1983), Sejrup (1987), (L.Olsen, unpubl.)
45, Vatnedalen, Rogaland	C14: 35,850 ±1180	T-2380	Palaeosol in subtill sed.	1	Blystad and Selsing (1988)
46, Elgane, Jæren, Rogaland	C14: 33,480 ±1520 C14: 34,820 ±1165	TUa-…	Shell, in reworked subtill sediments	1	Janocko et al. (1998)
47, Passebekk, Telemark	C14: 28,600 ±300	UtC 6044	BPR in reworked subtill sediments	2	Olsen et al. (2001a)
48, Rundhaugen, Telemark	C14: 32,000 ± …	UtC …	BPR in reworked subtill sediments*****	1	Roaldset (1980), Olsen and Grøsfjeld (1999)
49, Herlandsdalen, Telemark	C14: 28,300 ±240 C14: 32,000 ±300	UtC 4729 UtC 4728	BPR in reworked subtill sediments	2	Olsen et al. (2001a)

*Table B10: LGM interval oscillation sites (with sediments and tills). *Ages are shown in ^{14}C yr BP ±1std for ^{14}C dates and in cal yr BP for other dates. **BPR= Bulk plant remains. Average number of younger (overlying) stadials/glacial units is c. 1.16 (n=42), or c. 1.00 (n=32) with all sites which have no dates in the interval 15–21 ^{14}C-ka excluded (10 of 42 sites).*

Locality	Material	Dating*	Lab.refr.	Comments	Overlying glacial units	References
1, Leirelva, Finnmark	BPR** in subtill sediments	C14:17,110–17,290 and 18,680 ±170	UtC 1800,1799 and 3460	Probably LGM 1–2 oscillation, also TL and OSL	1	Olsen et al. (1996a), Olsen et al. (2001a)
2, Skjellbekken, Finnmark	BPR in subtill sediments	C14: 25,860 ±280	UtC 4040	Pre-LGM 1–LGM oscillation	2	Olsen et al. (2001a)
3, Arnelund, Troms	BPR in subtill sediments	C14: 16,580 ±100	UtC 12691	LGM 1–LGM 2 oscillation	1	(L.Olsen, unpubl.)
4, Andøya and Andfjorden sites	Ice marginal deposits, marine/lake sediments	C14 (several dates) range 17,940–21,800	TUa- …	LGM 1, Andøya interst., LGM 2, and younger	1	Vorren (1978), et al.(1988), Møller et al. (1992), Alm (1993)
5, Kjøpsvik, cave Nordland	Bone in sed. underlying ice-dammed sediments	C14: 20,110 ±250 C14: 20,210 ±130	TUa-436 TUa-488	LGM 1–LGM 2 oscillation	1	Nese and Lauritzen (1996), Olsen et al. (2001a)
6, Urdalen, Nordland	BPR in reworked sediment between till beds	C14: 20,470 ±110	UtC 8458	Pre-LGM 1–LGM 1– LGM 2 oscillations	1	Olsen et al. (2001a)
7, Meløya, Nordland	BPR in subtill sediments	C14: 17,700 ±80	UtC 8456	LGM 1–LGM 2 oscillation	1	Olsen et al. (2001a)
8, Kjeldal, Nordland	Succession of tills/ subtill sed. with resed. org.	C14: 18,880 ±100	UtC 8457	Ørnes interst.–LGM 1–Kjeldal interst.–LGM 2	1	Olsen (2002)
9, Bogneset, Nordland	Succession of tills/ subtill sed. with resed. org.	C14: 20,880 ±130	UtC 10109	Ørnes interst.–LGM 1–Kjeldal interst.–LGM 2	1	Olsen (2002)
10, Fiskelauselva Nordland	Succession of tills/ subtill sed. with resed. org.	C14: 19,500 ±200	UtC 2216	Hattfjelldal interst. II– LGM 1–LGM 2 oscillation	1	Olsen (1997)
11, Hattfjelldal Nordland	Succession of tills/ sub till sed. with resed. org.	C14: 23,500 ±240	UtC 4809	Hattfjelldal interst. II–LGM 1–LGM 2 oscillation	2	Olsen (1997)
12, Langstrandbakken Nord-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 18,700 ±500	UtC 5974	Andøya–Kjeldal interst.–LGM 2 ice advance	1	Olsen et al. (2001a)
13, Namsen, Nord-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 16,110 ±120 C14: 18,580 ±140	UtC 4811 UtC 4812	Andøya/ Kjeldal/ Trofors interst.–LGM 2 advance	1	Olsen et al. (2001a)
14, Øyvatnet, Nord-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 19,340 ±150	UtC 4800	Andøya/ Kjeldal/ Trofors interst.–LGM 2 advance	1	Olsen et al. (2001a)
15, Gartland, Nord-Trøndelag	BPR in reworked sediment in till	C14: 16,250 ±190	UtC 4871	Trofors interstadial–LGM 2 ice advance	1	Olsen et al. (2001a)
16, Sitter, Nord-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 21,150 ±130	UtC 4717	Andøya–Kjeldal interst.–LGM 2 ice advance	2	Olsen et al. (2001a)
17, Ø. Tverråga, Nord-Trøndelag	BPR in subtill sediment	C14: 17,830 ±190	UtC 3464	Trofors interstadial–LGM 2 ice advance	1	Olsen et al. (2001a)
18, Blåfjellelva I–II Nord-Trøndelag	BPR in sediments between till beds	C14: 19,710 ±110 C14: 20,040 ±100	UtC 5565 UtC 5566	LGM 1–Trofors interst.–LGM 2 ice advance	1	Olsen et al. (2001a)
19, Hangran, Sør-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 24,550 ±240	UtC 14602	Hattfjelldal II /Ørnes interst.– LGM 1/LGM 2 advance(s)	2	Olsen et al. (2007)
20, Renåa, Sør-Trøndelag	BPR in subtill sed., with trace of marine org.	C14: 16,850 ±90 C14: 19,880 ±160	UtC 5550 UtC 5551	LGM 1(?)–Trofors interst.–LGM 2 ice advance	1	Olsen et al. (2001a)
21A, Langsmoen, Sør-Trøndelag	Subtill sed., with resed. marine microfossils	OSL: 22,300 ±1700 (mean age, n=7)		Trofors interstadial– LGM 2 ice advance	1	Olsen et al. (2002), Johnsen et al. (2012)
21B, Flora, Sør-Trøndelag	BPR in subtill/ intertill sediments	C14: 17,550 (mean age, n=8)	UtC …	LGM 1–Trofors interstadial, followed by LGM 2 advance	1	Olsen et al. (2001a)
22, Stærneset, Sør-Trøndelag	BPR in sediments between till beds	C14: 18,820 ±110	UtC 5555	Hattfjelldal II–LGM 1–Trofors interstadial–LGM 2	1	Olsen et al. (2001a)
23, Surna, Møre and Romsdal	BPR in subtill sediment	C14: 19,090 ±100	UtC 10110	Trofors interstadial– LGM 2 ice advance	1	(L.Olsen, unpubl.)
24, Grytdal, Sør-Trøndelag	BPR in sediments between till beds	C14: 18,970 ±150	UtC 6040	Hattfjelldal II–LGM 1–Trofors interstadial–LGM 2	1	Olsen et al. (2001a)
25, Hauka, Sør-Trøndelag	BPR in subtill sediment	C14: 23,770 ±240 C14: 23,910 ±220	UtC 14605 UtC 14606	Hattfjelldal interstadial II–LGM 1/LGM 2 advance(s)	2	(L.Olsen, unpubl.)
26, Kvikneskogen Hedmark	BPR in sediments between till beds	C14: 16,660 ±100	UtC 12701	LGM 1(?)–Trofors interst.–LGM 2 ice advance	1	(L.Olsen, unpubl.)
27, Gamlemsveten Møre and Romsdal	BPR in soil overlain by blocks in block field	C14: 19,900 ±210	T-4384	Probably Andøya/Kjeldal/ Trofors interstadial	0	Mangerud et al. (1981a), (J.Mangerud, pers. comm. 1981)

Locality	Material	Dating*	Lab.refr.	Comments	Overlying glacial units	References
28, Hamnsundheller-en, Møre and R.	Bones in sed. between ice-dammed sediments	C14: 24,387 ±960 C14: 24,555 ±675	TUa-806 I TUa-806	Hamnsund/Ørnes/Hattfj.II int.st.–LGM 1/2 advance(s)	1	Valen et al. (1996)
29, Folldal, Hedmark	BPR in sediments between till beds	C14: 23,260 ±160	UtC 4710	Hattfjelldal interstadial II–LGM 1/LGM 2 advance(s)	2	Olsen et al. (2001a)
30, Kollsete, Sogn and Fjordane	BPR in subtill sediment	C14: 22,490 ±180	UtC 6046	Hamnsund/Ørnes/Hattfj.II int.st.–LGM 1/2 advance(s)	1	Aa and Sønstegaard (1997), Olsen et al. (2001a)
31, Stampesletta, Oppland	BPR in subtill/ intertill sediments	C14: c. 16,000 ± …	TUa-…	Trofors interstadial– LGM 2 ice advance	1	Olsen et al. (2001a)
32, Mesna, Oppland	BPR in subtill/ intertill sediments	C14: 16,030 ±100	UtC 6041	Trofors interstadial– LGM 2 ice advance	1	Olsen et al. (2001a)
33, Dokka, Oppland	BPR in palaeosol/subtill sediments	C14: 18,900 ±200	UtC 2218	Trofors interstadial– LGM 2 ice advance	1	Olsen (1998), Olsen et al. (2001a)
34, Bø, Karmøy, Rogaland	Shells from sediments in and between tills	C14: 18,770 ±160 (Mya truncata)	TUa-4519	LGM 1–Andøya/Kjeldal interst.–LGM 2 advance	1	Olsen and Bergstrøm (2007)
35, Foss-Eikeland Rogaland	BPR in subtill sediment	C14: 19,895 (mean age, n=4)	UtC …	Andøya/ Kjeldal interst.–LGM 2 ice advance	1	Olsen et al. (2004), Olsen and Bergstrøm (2007)
36, Haugastaulen, Hardangervidda	Subtill sand	OSL: 25,000 ±3000 OSL: 26,000 ±2000	R-33506 R-33504	Ice free conditions, and LGM 1/2 ice advance(s)	2	Haug (2005)
37, Passebekk, Telemark	BPR in subtill sediment	C14: 21,000 ±400	UtC 5987	Andøya/Trofors interstadial– LGM 2 ice advance	1	Olsen et al. (2001a)
38, Herlandsdalen, Telemark	BPR in subtill sediments	C14: 23,250 ±170	UtC 6045	Hattfjelldal II followed by LGM 1/LGM 2 advance(s)	2	Olsen et al. (2001a)
39, Skjeberg, Østfold	BPR in subtill sediments	C14: 16,770 ±190 C14: 19,480 ±200	UtC 1802 UtC 1801	Andøya/Kjeldal interstadial– LGM 2 ice advance	1	Olsen et al. (2001a)
40, Braarød, Østfold	BPR in subtill sediments	C14: 19,260 ±130	UtC 12705	Andøya/Kjeldal interstadial– LGM 2 ice advance	1	Olsen et al. (2004), Olsen and Bergstrøm (2007)
41, Korterød, Østfold	BPR in subtill sediments	C14: 16,680 ±140	UtC 12707	Andøya/Kjeldal interstadial– LGM 2 ice advance	1	Olsen et al. (2004), Olsen and Bergstrøm (2007)
42, Børsåsen, Østfold	BPR in subtill sediments	C14: 15,110 ±90	UtC 12703	Ice free conditions and LGM 2 or younger ice advance	1	Olsen et al. (2004), Olsen and Bergstrøm (2007)

Table B11: Post LGM–pre-Older Dryas oscillation sites (with sediments and tills).

Post LGM–pre Older Dryas localities. Mean age of oldest post-LGM oscillation on land = 17.5 cal ka (14.5 ^{14}C ka) (N=18)		
Locality	**^{14}C age (ka)**	**Comments**
1, Komagelva	15.2 (N=4)	Subtill position, plant remains
2, Vuoddasjavri	13.8	Subtill position, plant remains, possible young C contamination
3, Andøya	c. 16	Glaciomarine and freshwater sed., plant remains
4, Hinnøya	13.9 and 14.0	Subtill position, dislocated, plant remains, traces of marine fossils
5, Domåsen	14.1 and 14.7	Subtill position, possible subgl. resed./or young C contamination
6, Myrvang	14.4	Subtill position, plant remains, traces of marine fossils
7, Hangran	14.7	Subtill position, plant remains, traces of marine fossils
8, Rønningen	13.9	Subtill position, plant remains, possible young C contamination
9, Flora	14.7	Subtill position, plant remains, possible young C contamination
10, Grytdal	13.5	Subtill position, plant remains, possible young C contamination
11, Bø	15.3	Subtill position, marine shell (Mya truncata)
12, Lista	c. 13.5–14.0	End moraine (Lista Substage)/Ice stream 'lateral shear moraine', estimated age
13, Børsåsen	15.1	Subtill position, plant remains

Localities:	Geography:	References:
1, 2	North Norway	Olsen et al. (1996a)
3	North Norway	Alm (1993), Vorren et al. (1999)
4	North Norway	Olsen et al. (2001a)
5, 6, 9, 10	Mid Norway	Olsen et al. (2001a)
7	Mid Norway	(L. Olsen, unpublished 2006)
8	Mid Norway	Olsen et al. (2007)
11	South Norway	Olsen and Bergstrøm (2007)
12	South Norway	Andersen (1960, 2000), (NGU unpubl.)
13	South Norway	Olsen et al. (2004)

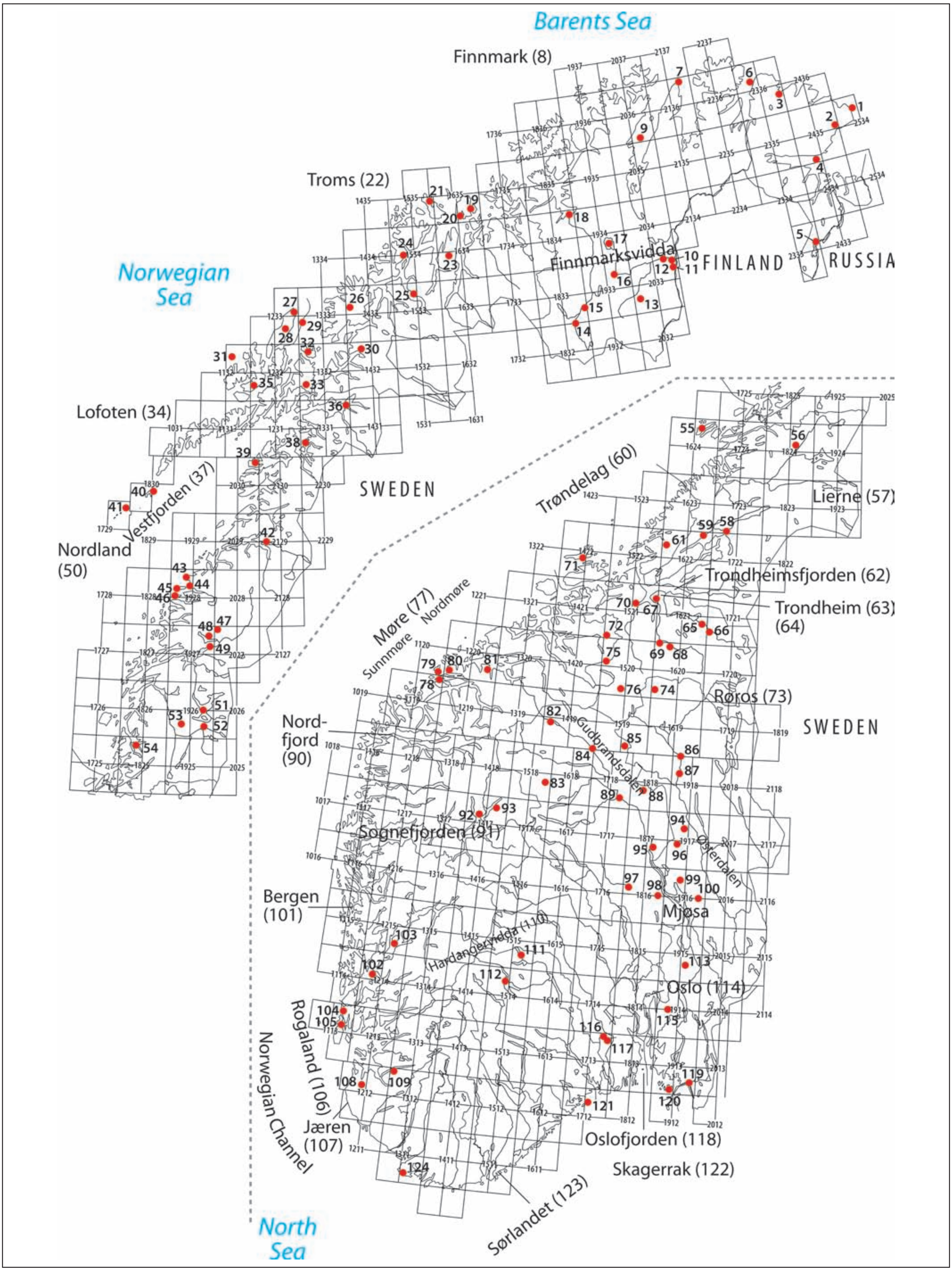


Figure B1: Location of sites and places listed by numbers and names in Table B1.

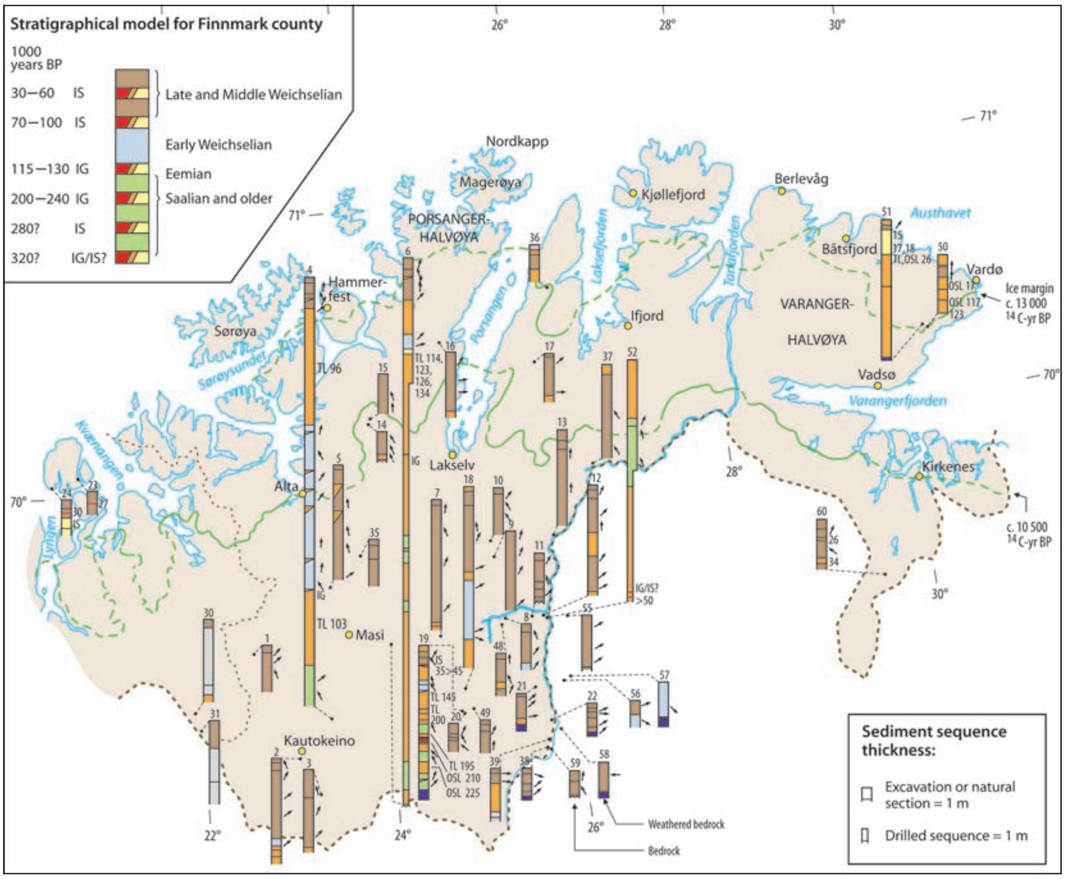


Figure B2: Part of the Quaternary stratigraphical map of Finnmark (Olsen 1993b). The logs from sites 4, 6, 52 and 19 here are shown as nos. 1, 2, 4 and 6, respectively, in Figure 13 (main text).

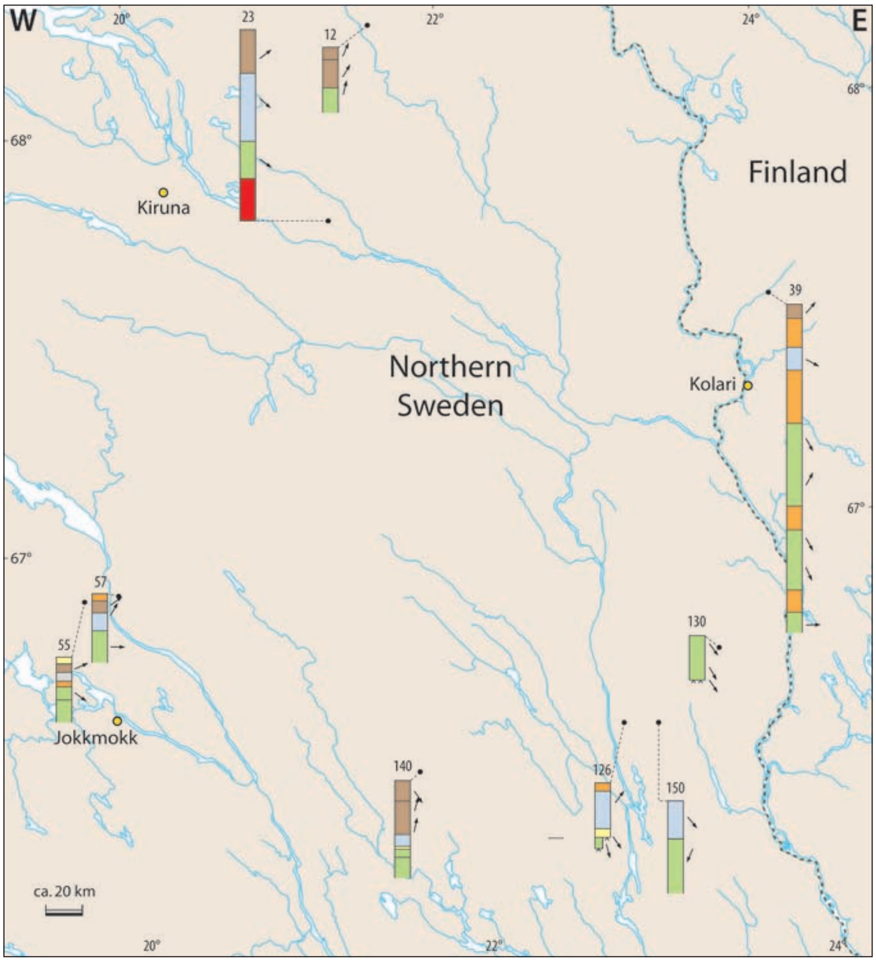
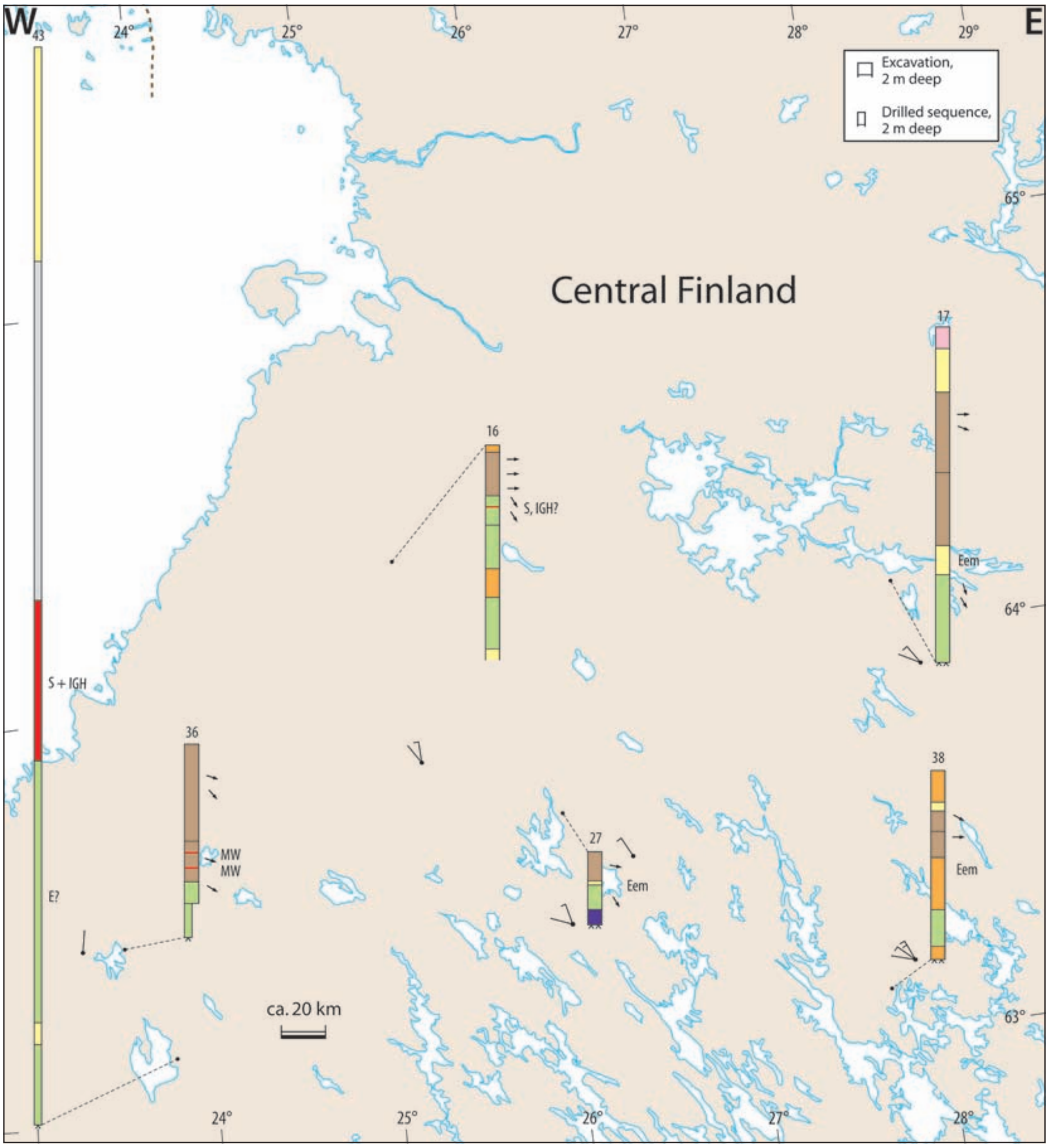


Figure B3: Logs from selected sites in the central part (mainly Sweden) of the Quaternary stratigraphical map of northern Fennoscandia (The Nordkalott Project 1986).

Figure B4: Logs from selected sites in the eastern part (Finland) of the Quaternary stratigraphical map of central Fennoscandia (The Mid-Norden Project 1999).



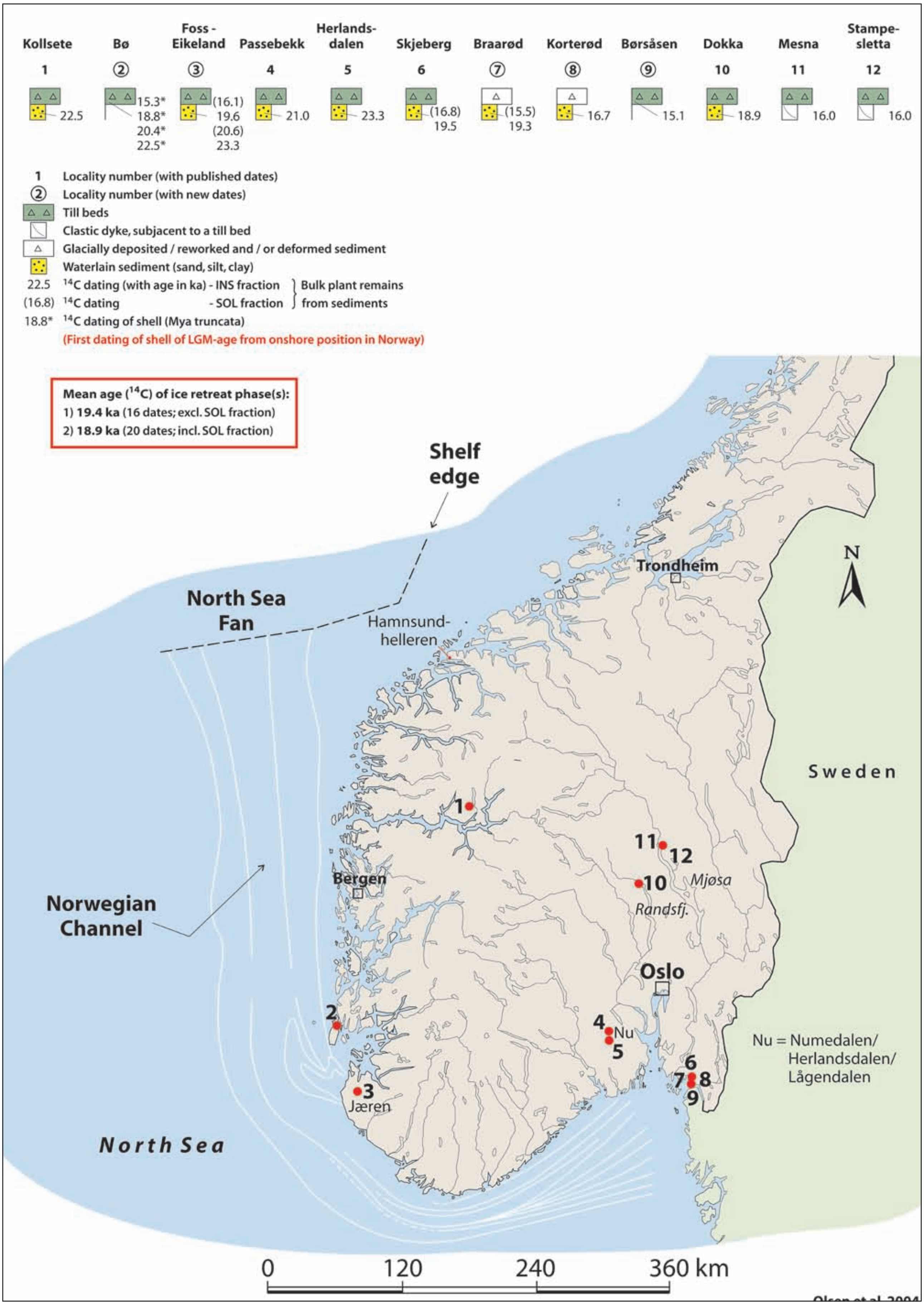


Figure B5: LGM sites on land adjacent to the Norwegian Channel along the coast of southern Norway. After Olsen et al. (2004).

Neotectonics, seismicity and contemporary stress field in Norway – mechanisms and implications

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Neotectonics in Norway are characterised by: 1) geological features: two documented postglacial faults in northern Norway; Neogene doming of sedimentary depocentres in the Vøring area. 2) seismicity: enhanced earthquake activity along the coastal areas of northern, western and southeastern Norway; palaeoseismic events in western and northern Norway; present-day seismicity along the Stuoragurra postglacial fault indicates that the fault is active at depth. 3) rock stress: local deviations from a general NW–SE-oriented compressional in situ rock stress; areas with observed extension from fault-plane solutions in western and northern Norway. 4) uplift: increasing present-day uplift from west to east with the highest values in Trøndelag and eastern Norway (4 mm yr^{-1}); Neogene long-term uplift of western and northern Norway as indicated by raised pre-Weichselian sediments and coastal caves; an active area of extension and subsidence in the outer Ranafjorden area. These neotectonic features are likely to be mostly related to gravitational effects of excess mass along the Mohns Ridge, within the Iceland Plateau and the southern Scandinavian mountains, to Pliocene/Pleistocene sedimentary loading/unloading, and to postglacial rebound. A major seismic pulse most likely accompanied each of the deglaciations following the multiple glaciation cycles in mainland Fennoscandia during the last 600,000 years. Seismic pumping associated with these glaciation cycles may have facilitated fluid and gas leakage from organic-rich sediments and reservoirs through gas chimneys, ultimately forming pockmarks on the sea floor. This mechanism could also have contributed to the concentration and pumping of hydrocarbons from their source rocks to reservoir formations. Pressure decrease associated with removal of sedimentary overburden on the Norwegian shelf has caused expansion of gas and resulted in expulsion of oil from the traps. Where uplift and tilting resulted in local extension, seal breaching and spillage have also occurred. Future rock avalanches and landslides, triggered by earthquakes, could generate tsunamis in fjords and lakes and constitute the greatest seismic hazard to society in Norway. Our understanding of neotectonic activity is consequently important for the evaluation of hazard and risk related to rock-slope instability.

Olesen, O., Bungum, H., Dehls, J., Lindholm, C., Pascal, C. and Roberts, D. (2013) Neotectonics, seismicity and contemporary stress field in Norway – mechanisms and implications. *In* Olsen, L., Fredin, O. and Olesen, O. (eds.) *Quaternary Geology of Norway*, Geological Survey of Norway Special Publication, **13**, pp. 145–174.

Introduction

Over the last thirty years, through studies of neotectonic phenomena, it has become evident that the present-day Baltic Shield is not the uniformly quiet, stable, continental-crustal area that was earlier commonly assumed. In Norway, and northern Fennoscandia as a whole, detailed seismotectonic investigations, recordings of Late Quaternary faults, stress measurements and observations of stress-release features have all indicated that neotectonic movements have been, and still are, quite significant (Kujansuu 1964, Lagerbäck 1979, 1990, Olesen 1988, Bungum 1989, Slunga 1989, Talbot and Slunga 1989, Roberts 1991, 2000, Olesen et al. 1992a, 1995, 2004, Myrvang 1993, Bungum and Lindholm 1997, Muir Wood 1989a, 2000, Stewart et al. 2000, Roberts and Myrvang 2004, Pascal et al. 2005a,b, 2006). Monitoring of seismicity in the adjacent continental shelf, together with data from borehole breakouts, has also greatly increased our knowledge of the contemporary, regional stress regime (Bungum et al. 1991, Gölke and Brudy 1996, Hicks et al. 2000a, Byrkjeland et al. 2000), a factor of no mean importance for the offshore petroleum industry. Neotectonics, and the potential hazards associated with such crustal motions, thus constitute an important component of the Quaternary geology of Norway.

Our approach to neotectonic studies follows the definition of neotectonics as given by the International Association for Quaternary Research (INQUA); "Any earth movement or deformations of the geodetic reference level, their mechanisms, their geological origin (however old they may be), their implications for various practical purposes and their future extrapolations" (INQUA 1982). In Norway, the first known report of neotectonic activity is that of Morsing (1757), and over the last 250 years the number of such reports has increased steadily and has now reached more than 80 (Olesen et al. 2004). Additional reports deal with three shallow earthquake swarms along the coast of Nordland (Bungum and Husebye 1979, Bungum et al. 1979, Atakan et al. 1994, Hicks et al. 2000b) and four separate swarms on Svalbard (Bungum et al. 1982, Mitchell et al. 1990, Pirli et al. 2010) that could, dependent on definition, be added to the list (Figure 1). Almost 80% of the reports were published after 1980. Twenty of the claims are situated in the offshore area and more than 60 are located on mainland Norway.

A first coordinated attempt to assess the status and many facets of neotectonic activity in Norway came with the 'Neotectonics in Norway' (NEONOR) project during the years 1997–2000. This aimed at investigating neotectonic phenomena through an integrated approach including structural bedrock mapping, monitoring of microseismicity, recording of stress-release features, study of aerial photographs, trenching, drilling, ^{14}C dating, geodetic levelling and ground-penetrating radar profiling (Dehls et al. 2000a, Fjeldskaar et al. 2000, Hicks et al. 2000a, b, Roberts 2000, Olesen et al. 2004, Rise et al. 2004). Seismic surveying (including available 3D data) and multibeam echo-sounding techniques were used to examine possible offshore postglacial faulting. The shallow parts of eight seismic 3D cubes (located in seismically active areas) were studied to try to locate potential Quaternary deformation features. Results from rock-avalanche hazard projects in Troms and western Norway (Geological Survey of Norway [NGU]) and the 'Seabed Project' (NORSAR/NGI/UiO/SINTEF) were also included in this major assessment of neotectonic activity on the Norwegian mainland and continental shelf.

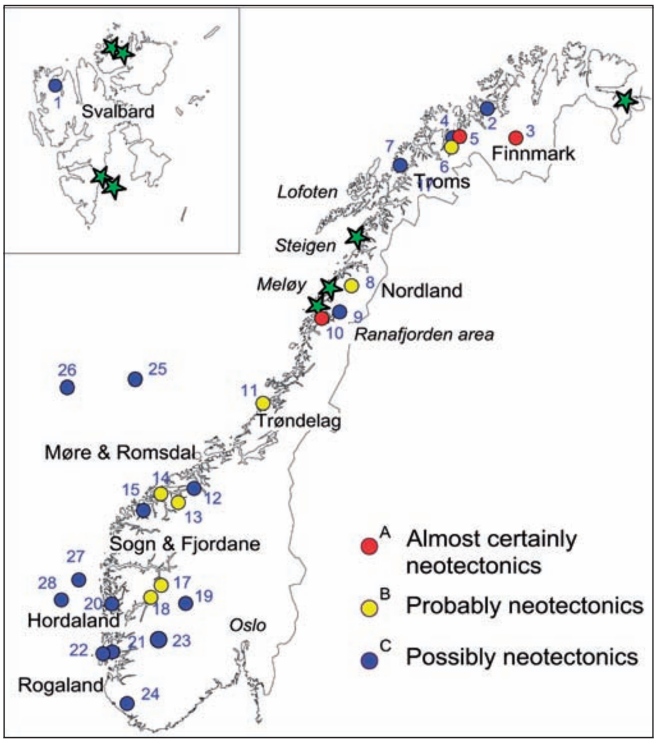


Figure 1. Locations of 28 neotectonic claims that have been classified as A, B and C. Location No. 16 covers the coastal area between Sogn & Fjordane and Lofoten and is not shown on the map. Reports with grades D and E are not included, but are shown on a similar map by Olesen et al. (2004). Green stars denote shallow earthquake swarms (Bungum and Husebye 1979, Bungum et al. 1979, 1982, Mitchell et al. 1990, Atakan et al. 1994, Hicks et al. 2000b, Pirli et al. 2010). The numbers refer to information on the neotectonic claims listed in the Appendix.

In this paper we summarise our current knowledge and understanding of neotectonics in Norway by grouping the reports and data into four categories, namely: postglacial faulting, postglacial and contemporary uplift, seismicity, and the contemporary stress field.

Postglacial faulting

Onshore

Two postglacial faults have been documented on mainland Norway (Olesen 1988, Tolgensbakk and Sollid 1988). The NE–SW-oriented, reverse Stuoragurra Fault (Olesen 1988, Muir Wood 1989a, Olesen et al. 1992a,b, Dehls et al. 2000a) in western Finnmark constitutes the Norwegian part of the postglacial Lapland Fault Province (Figures 2–5, Table 1). The fault consists of numerous segments within a 80 km-long and 2 km-wide zone and has a maximum scarp height of 7 m (Figures 3a and 6). The dip of the fault is approximately 55° to the southeast near the surface (Figure 6a), A total of three percussion drillholes and one core drilling down to a depth of 135 m are located along a profile perpendicular to the Stuoragurra postglacial fault (Olesen et al. 1992a,b, Roberts et al. 1997, Dehls et al. 1999, Kukkonen et al. 2010). The drillholes revealed that the postglacial fault at a depth of *c.* 50 m has a dip of *c.* 40° to the southeast and consists of several thin (a few cm thick) zones of clay minerals within a 1.5 m-thick interval of fractured quartzite (Olesen et al. 1992a,b, Roberts et al. 1997). The clay zones consist of kaolinite, vermiculite, smectite, goethite and chlorite, and most likely represent a weathered fault gouge (Åm 1994). Several 2–3 m-thick zones of lithified breccia within a 25 m-wide interval reveal that the postglacial fault occurs within an old zone of weakness partly coinciding with the margins of deformed Palaeoproterozoic albite diabases. Magnetic modelling of the albite diabase in the vicinity of the drillholes shows a dip of *c.* 40° to the southeast (Olesen et al. 1992a,b) consistent with the results from the drilling. Resistivity and refraction seismic profiling show both low resistivity (900 ohmm) and low seismic velocity (3800 m s^{-1}) and indicate a high degree of fracturing.

Focal mechanism solutions for five earthquakes recorded along or close to the Stuoragurra Fault and observation of stress-release features in Finnmark (Roberts 2000, Pascal et al. 2005a) have indicated that the maximum principal compressive stress, S_{Hmax}, is oriented approximately NW–SE. The individual focal mechanisms (Bungum and Lindholm 1997) were poorly constrained and were located southeast of the Stuoragurra surface expression at shallow depths. The reverse/oblique nodal planes were oriented so that one plane could be associated with the fault strike for all events; however, σ_H varied from N–S to E–W (averaging to NW–SE).

Olesen (1988) and Muir Wood (1989a) noted that both the Pärvie and Stuoragurra faults occur along a physiographic border. The mountainous area to the northwest has an average higher elevation than the area to the southeast. The ice was consequently thickest in the southeastern area. This would have involved more depression during the period of maximum glaciation and consequently a greater contribution to the subsequent postglacial stress regime. The differential loading of ice across a prestressed zone of weakness might consequently be sufficient to have caused reactivation of the old zone, and so produce a fault scarp.

This model, however, will not explain the other postglacial faults in Fennoscandia. Muir Wood (1993, 2000) suggested that interference between polarised tectonic (ridge push) and radial deglaciation strain fields produce alternating quadrants of enhanced seismicity and aseismic regions around rebound domes and former peripheral forebulges. He argued that the observed postglacial faults occur within such a seismic quadrant

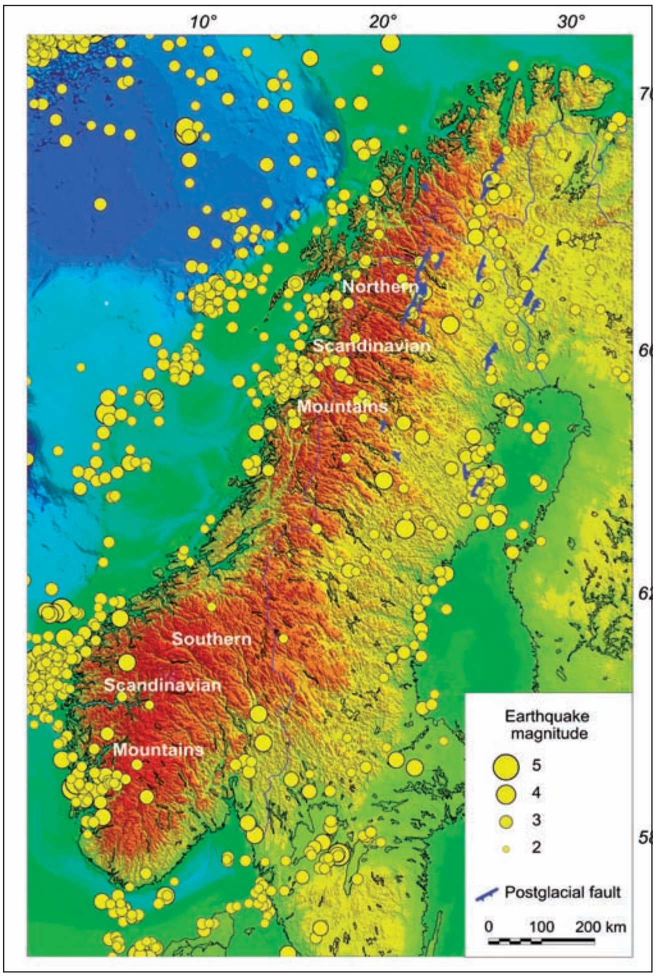


Figure 2. Earthquakes during the period 1980–2011 and postglacial faults in Fennoscandia (modified from Dehls et al. 2000b, Olesen et al. 2004, Lagerbäck and Sundh 2008 and Bungum et al. 2000, 2010). The Norwegian National Seismological Network at the University of Bergen is the source of the earthquake data in Norway, Svalbard and NE Atlantic. Data on the other earthquakes in Denmark, Finland and Sweden are downloaded from the web pages of the Institute of Seismology at the University of Helsinki; http://www.seismo.helsinki.fi/english/bulletins/index.html. We have established a lower threshold at magnitude 2.5 to reduce the probability of contamination by explosives. The size of the earthquake symbols increases with rising magnitude. The postglacial faults occur in areas with increased seismicity.

in northern Fennoscandia where ridge-push stress and rebound stress are superimposed.

The 2 km-long and NW–SE-striking Nordmannvikdalen fault (Figure 3b, Tolgensbakk and Sollid 1988, Dehls et al. 2000a) near Kåfjord in northern Troms is a normal fault trending perpendicular to the NE–SW-trending, postglacial reverse faults in northern Fennoscandia. The kinematics of the fault still fit the regional stress pattern and may relate to local relief effects allowing release along the NW–SE trend. The Nordmannvikdalen fault may also be considerably longer but its full extent is difficult to estimate because of missing overburden along the possible extensions to the southeast and northwest.

The NNE–SSW-trending and reverse Berill Fault (site 13 in Figure 1 and Appendix, Figure 3c and 3d) occurs in Møre & Romsdal county in southern Norway (Anda et al. 2002)

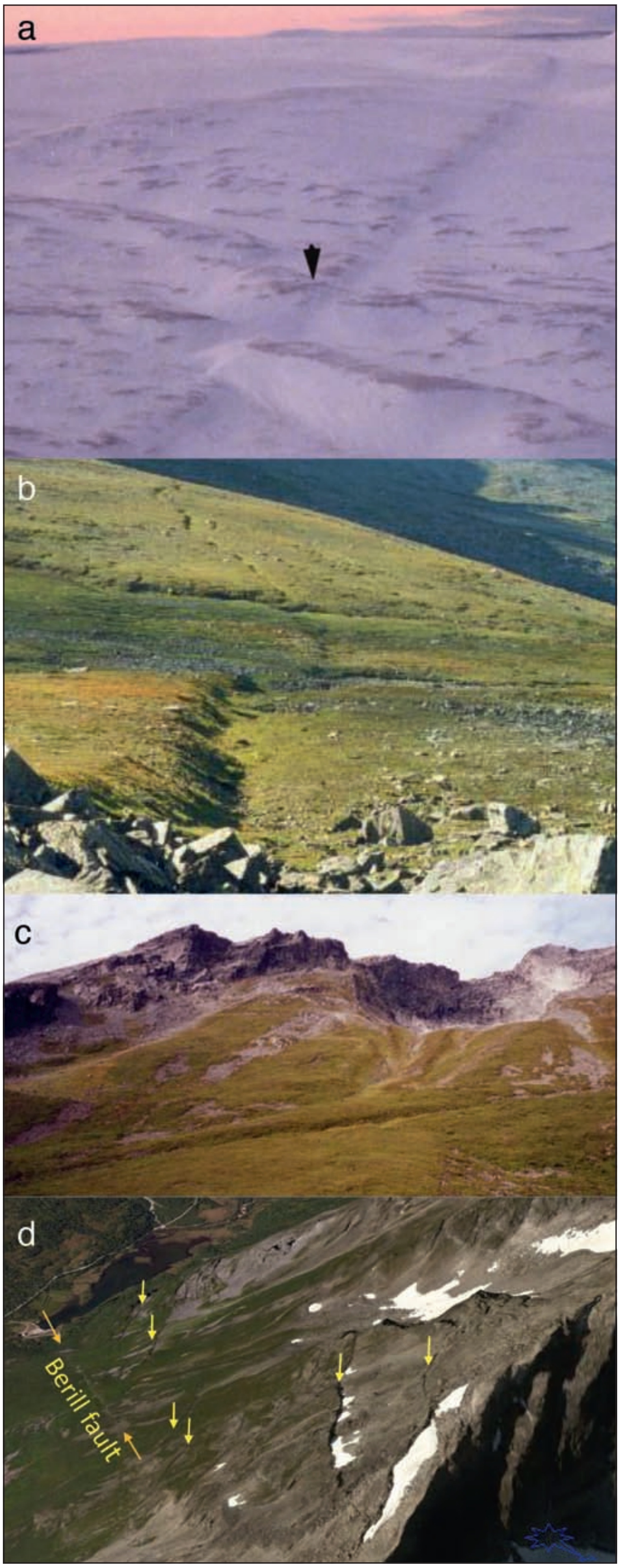


Figure 3. Two postglacial faults in Norway and a sackung structure previously classified as a tectonic fault. (a) Oblique aerial photograph of the Stuoragurra Fault (Location 3 in Figure 1 and fault 1 in Figure 5 and Table 1) as it crosses Finnmarksvidda at Stuoragurra, 15 km NNE of Masi (Olesen et al. 2004). The fault cuts through an esker (UTM 611400E, 7717300N). The intersection between the fault scarp and the esker is shown by the arrow. The fault was consequently formed after the deglaciation at approximately 9300 BP. (b) The Nordmannviken Fault (fault 2 in Figure 5 and Table 1) viewed from the southeast (Dehls et al. 2000a). The fault scarp runs parallel to the valley floor. The surface slope is at most 15–20° and it is therefore not likely that the fault scarp is due to gravitational sliding. (c) The NNE–SSW-trending and reverse Berill Fault (Anda et al. 2002) has previously been classified as a tectonic fault (Olesen et al. 2004). The length of the fault is minimum 2.5 km and the scarp height is 3 m and it dates most probably to the latter half of the Holocene. This reverse fault is located at the foot of the Middagstinden mountain and appears to be part of a sackung structure (see text). (d) Open clefts with upward-facing scarps (yellow arrows) along the Middagstinden mountain ridge occur in the hanging-wall block of the Berill Fault (located between the orange arrows) and are typical features of gravity-induced sackung structures. The image is produced using www.Norgei3D.no.

and has previously been classified as a tectonic fault (Olesen et al. 2004). The length of the fault is minimum 2.5 km and the scarp height is 3 m and it dates probably to the latter half of the Holocene. This reverse fault is located at the foot of the Middagstinden mountain and appears to be part of a sackung structure (Savage and Varnes 1987). The open clefts with upward-facing scarps along the mountain ridge in the hanging-wall block of the reverse fault (Figure 3d) are typical features of gravity-induced sackung structures. The low offset/length ratio (1:500) of the fault also points to a nontectonic origin. We have decided to remove the Berill Fault from the list of 'A – almost certainly neotectonics'. The structures may, however, have been triggered by adjacent large-magnitude earthquakes and the fault is therefore classified as 'B probably neotectonics' in Figure 1. For details on the remaining, probable and possible, neotectonic observations in Figure 1 (yellow and blue), readers are referred to Dehls et al. (2000a) and Olesen et al. (2004).

There seems to be an anomalously high number of rock avalanches in the vicinity of the Nordmannvikdalen fault suggesting a link between rock-slope failures and palaeoseismic events (Braathen et al. 2004, Osmundsen et al. 2009). The Nordmannvikdalen fault was most likely formed shortly after the deglaciation.

Dehls et al. (2000a) and Olesen et al. (2004) graded existing neotectonic reports into the following classes according to their reliability (Muir Wood 1993, Fenton 1994): (A) almost certainly neotectonics, (B) probably neotectonics, (C) possibly neotectonics, (D) probably not neotectonics and (E) very unlikely to be neotectonics. The most likely nature of the proposed neotectonic deformation was identified whenever possible and placed in the following categories; (1) tectonic faults, (2) gravity-induced faults, (3) erosional phenomena, (4) overburden draping of bedrock features, (5) differential compaction, (6) shallow, superficial stress-release features, (7) inconsistent shoreline correlation, (8) unstable benchmarks and levelling errors.

Critical evaluation of more than 60 neotectonic claims in mainland Norway and Svalbard has resulted in three claims of grade A and eight claims of grade B. The grade A claims include

*Table 1. Summary of properties of the documented postglacial faults in Finland, Norway and Sweden (modified from Olesen et al. 2004 and Lagerbäck and Sundh 2008). The major faults are NE–SW-trending reverse faults and occur within a 400 x 400 km area in northern Fennoscandia. The normal Nordmannvikdalen fault is a minor fault trending perpendicular to the reverse faults. The NW–SE-trending Storuman fault in northwestern Sweden may be an analogue to the Nordmannvikdalen fault but the sense of movement along the fault has not been studied yet (Lagerbäck and Sundh 2008). The NW–SE-trending Vaalajärvi fault in northern Finland has been removed from the table since it is most likely not postglacial (M. Paananen, pers. comm. 2007). The scarp height/length ratio is generally less than 0.001. The Merasjärvi Fault has a scarp height/length ratio of 0.002. *Moment magnitudes calculated from fault offset and length utilising formulas by Wells and Coppersmith (1994).*

No.	Fault	Country	Length (km)	Max. scarp height (m)	Height/ length ratio	Trend	Type	Moment magnitude	Comment	Reference
1.	Stuoragurra	Norway	80	7	0.0001	NE–SW	Reverse	7.3	Three separate sections	Olesen 1998
2.	Nordmannvik-dalen	Norway	2	1	0.0005	NW–SE	Normal	6.0		Tolgensbakk and Sollid 1988
3.	Suasselkä	Finland	48	5	0.0001	NE–SW	Reverse	7.0		Kujansuu 1964
4.	Pasmajärvi–Venejärvi	Finland	15	12	0.0008	NE–SW	Reverse	6.5	Two separate sections	Kujansuu 1964
5.	Pärvie	Sweden	155	13	0.0001	NE–SW	Reverse	7.6		Lundqvist and Lagerbäck 1976
6.	Lainio–Suijavaara	Sweden	55	*c.* 30	0.0005	NE–SW	Reverse	7.1		Lagerbäck 1979
7.	Merasjärvi	Sweden	9	18	0.002	NE–SW	Reverse	6.3	Possible extension of the Lainio–Suijavaara Fault	Lagerbäck 1979
8.	Pirttimys	Sweden	18	2	0.0001	NE–SW	Reverse	6.5		Lagerbäck 1979
9.	Lansjärv	Sweden	50	22	0.0004	NE–SW	Reverse	7.1		Lagerbäck 1979
10.	Burträsk–Röjnoret	Sweden	60	*c.* 10	0.0002	NE–SW N–S	Reverse	7.1	Two separate faults	Lagerbäck 1979
11.	Sorsele	Sweden	2	1.5–2	0.0009	NE–SW	Reverse	6.1		Ransed and Wahlroos 2007
12.	Storuman	Sweden	10	10	0.001	NW–SE	?	6.3	Several separate faults	Johansson and Ransed 2003

the two postglacial faults described in the sections above and an active area of extension and subsidence in the outer Ranafjorden area (Olesen et al. 2012b). The grade B claims include areas with secondary effects, probably triggered by large-magnitude earthquakes, such as liquefaction and semiliquefaction structures in the Flatanger (Nord-Trøndelag) and Rana (Nordland) areas, and gravitational spreading and faulting features (sackung) on Kvasshaugen in Beiarn (Nordland), Berill in Rauma and Øtrefjellet in Haram (Møre & Romsdal). A series of gravitational fault systems and large rock-slope failures in zones from Odda to Aurland (Hordaland and Sogn & Fjordane) and in northern Troms have also been classified as grade B. The gravitational spreading, gravitational faults and large-scale rock avalanches are obviously caused by gravity collapse, but their spatial occurrence and the relatively gentle slopes associated with some of the features indicate that another mechanism assisted in triggering these events (Anda et al. 2002, Braathen et al. 2004, Blikra et al. 2006). The most likely cause is strong ground shaking from large-magnitude earthquakes. Two examples of collapse structures (in Haram and Ulvik) occur in gently sloping terrain and were probably not induced by gravity alone. The Tjellefonna and Silset rock avalanches in 1756 in the Møre & Romsdal county were possibly caused by an earthquake (Morsing 1757) and are therefore classified as a C claim (possibly neotectonics) in the present study.

A majority of the neotectonic claims can consequently be attributed to causes other than tectonic (Olesen et al. 2004). Gravity-induced sliding and glacial erosion along pre-existing faults and fractures were the dominant agents responsible for forming the geomorphological features that were earlier claimed to be of neotectonic origin. Ice-plucking features may, however, be indirectly related to neotectonics. Bell and Eisbacher (1995) showed that moving glaciers in the Canadian Cordillera tend to pluck bedrock along extensional fractures parallel to the

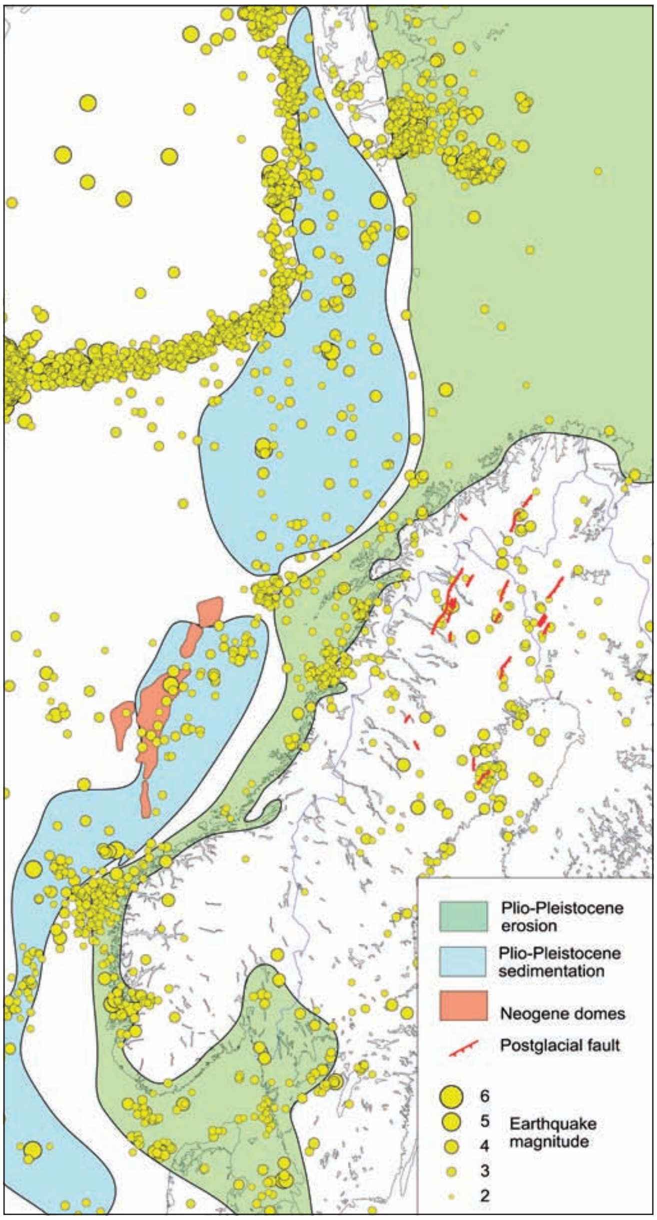


Figure 4. Earthquakes (1980–2011), postglacial faults, Neogene domes and areas of interpreted Pliocene/Pleistocene deposition and erosion along the Norwegian continental margin (modified from Blystad et al. 1995, Riis 1996, Lidmar-Bergström et al. 1999, Bungum et al. 2000, 2010, Dehls et al. 2000b, Lagerbäck and Sundh 2008, Kukkonen et al. 2010). The areas of Pliocene/Pleistocene sedimentation and erosion coincide with present-day seismicity, indicating that recent loading/unloading is causing flexuring and faulting in the lithosphere. The erosion of the central and southwestern Barents Sea may be older than the erosion of the Svalbard region and the coastal areas of northern, western and southeastern Norway since the seismicity of the former area is low. Focal-plane solutions (Dehls et al. 2000b, Hicks et al. 2000a) indicate the dominating compressional events in the areas with loading, whereas the regions with recent unloading have predominantly extensional or strike-slip events.

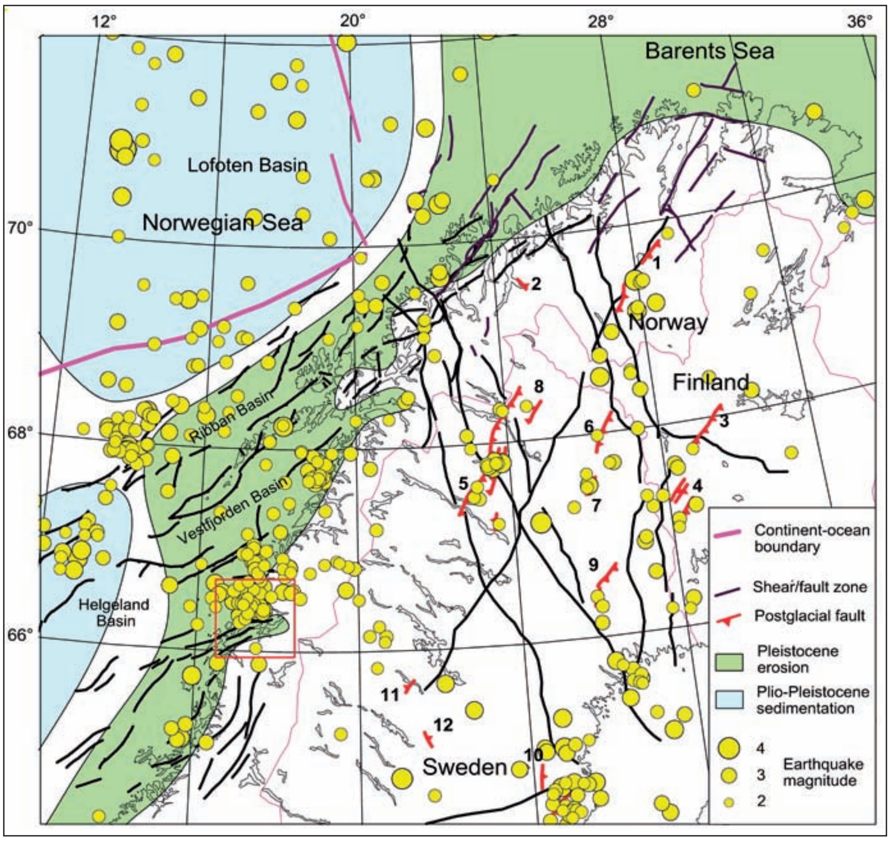
direction of maximum horizontal stress. An inland glacier could, in a similar way, cause a higher degree of bedrock plucking by basal glacier shear along favourably oriented fractures in areas with highly anisotropic rock stress.

The highest numbers of neotectonic claims have been reported from Rogaland, Hordaland and Nordland (Figure 1), but no postglacial faults have, up until now, been documented in these areas. Helle et al. (2007) made a new review of neotectonic reports from the former two counties. They emphasised the observed deviations from the general pattern in the Younger Dryas maximum highstand shoreline as indications of movements younger than *c.* 10,500 ^{14}C years BP. These deviations are in the order of 2–6 m and are mostly based on single observation points. Helle et al. (2007) were not able to relate these anomalous locations to any nearby postglacial faults. There are, however, indications of postglacial faulting on high-resolution, multibeam, echo-sounding data to the west of Bokn (Figs. 5.2 and 5.3 in Rønning et al. 2006). This 3 km-long and NNE–SSW-trending fault occurs along a line that may constitute an extension of Skjoldafjord to the southwest. There is unfortunately no high-quality bathymetry in the Skjoldafjord area that could be utilised to link the Bokn fault to the observed offset in the Yrkje area on the eastern shore of Skjoldafjord. The scarp faces west, consistent with the eastern part of the Yrkje area being uplifted. The height of the scarp along the Bokn fault varies between 0 and 60 m and this variation is too high to be related to faulting along this short fault segment (see criteria in Fenton 1994, Muir Wood 1993, Olesen et al. 2004). The postglacial faulting could, however, be superimposed on a pre-existing erosional scarp. The postglacial scarp can locally be covered by marine clay due to variability of the currents in the area.

Helle et al. (2007) also referred to observed apparent offsets of sediments on seismic profiles in Hardangerfjorden (Hoel 1992) as indications of postglacial faults. An interpretation of modern seismics and multibeam, echo-sounding surveys for the planning of a subsea power line in the Hardangerfjorden has revealed that the seafloor offsets are related to submarine landslides (Eriksson et al. 2011, Oddvar Longva, pers. comm. 2012).We therefore do not regard these reports as compelling evidence of postglacial faulting and have consequently graded them as D (probably not neotectonics). Modern, multibeam, echo-sounding surveying is an efficient method for scrutinising deep fjords and lakes in the potential neotectonic areas for postglacial faulting.

The Ranafjorden–Meløy area in Nordland is another area with numerous reports of neotectonic deformation (Figures 1 and 5). Olesen et al. (1995) suggested that the 50 km-long Båsmoen fault could be a candidate for a postglacial fault. They based their evidence on the observed escarpments facing NNW and an anomalous present-day uplift pattern along the fault. No conclusive evidence, however, has been found for postglacial movements along the fault (Olesen et al. 2004), although trenching has indicated a 40 cm offset along the fault in the Båsmoen area (Olsen 2000, Olesen et al. 2004). The observed seismicity (Hicks et al. 2000b) seems to occur along N–S-trending fractures and faults with pronounced escarpments in the Handnesøya–Sjona area (Figures 7 and 8 and Olesen et al. 1995). These scarps have been partly attributed to plucking effects by the moving inland ice. It is also intriguing that there

Figure 5. Earthquakes (M >2.5) during the period 1980–2011, postglacial faults and main basement structures in northern and central Fennoscandia. Details of the postglacial faults are shown in Table 1. The postglacial faults occur in areas with increased seismicity indicating that they are active at depth. The numbers adjacent to the faults refer to the numbers in Table 1. The interpreted basement structures (shear zones and detachments) from northern Fennoscandia are compiled from Henkel (1991), Olesen et al. (1990, 2002) and Osmundsen et al. (2002).



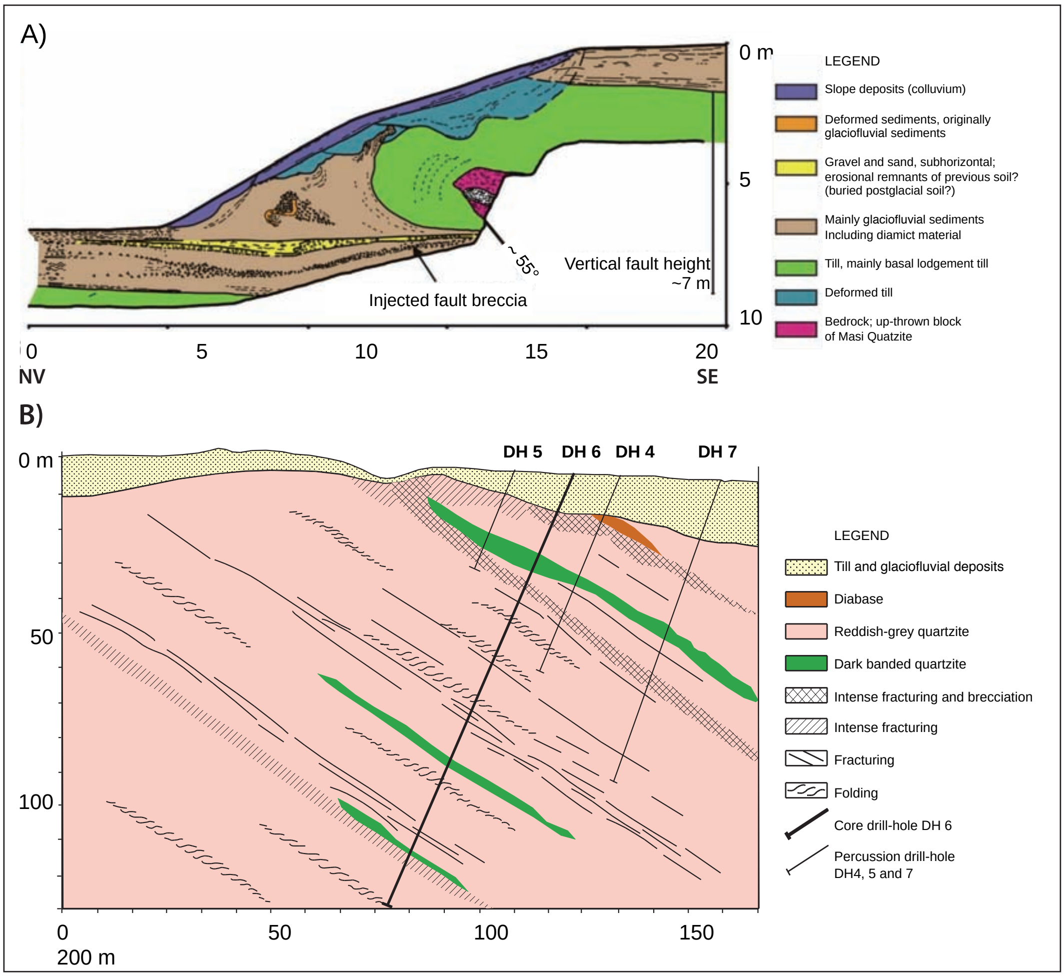


Figure 6. Stuoragurra Fault. (a) Folded Quaternary sequence consisting of basal till and glaciofluvial sediments above and in front of the up-thrown hanging-wall block of the Stuoragurra Fault. Note the fault breccia that has been injected into the glaciofluvial sediments most likely as a mixture of rock fragments and high-pressure groundwater (Modified from Dehls et al. 2000a). (b) Interpretation profile across the Stuoragurra Fault based on core (DH6) and percussion (DH4, 5 and 7) drilling and geophysical measurements (Olesen et al. 1992a,b, Roberts et al. 1997, Olesen et al. 2000, Kukkonen et al. 2010). The trench in (a) is located c. *100 m to the northeast of the profile in (b). See Figure 16 in Olesen et al. (1992b) for more details on the two locations.*

Figure 7. Several vertical fracture zones on Handnesøya (Olesen et al. 1994, 1995). The western blocks seem to be downfaulted. Part of the scarps could be the effect of subglacial plucking from the moving inland ice (Olesen et al. 2004). The two westernmost scarps appear to coincide with the linear seismicity clusters in Figure 8. Looking north from the quay in Nesna.

are no observed offsets of marine sediments on reflection-seismic profiles in the fjord to the north or the south of Handnesøya (Longva et al. 1998).

Several independent datasets in the outer Ranafjorden region indicate that the area is currently experiencing a regime of WNW–ESE extension (Figure 8). A six-station seismic network in this region during an 18 month period from July 1997 to January 1999 detected *c.* 300 earthquakes, many of them occurring as swarms. Fault-plane solutions indicate E–W extensional faulting. The outer Ranafjorden district is also the location of the largest earthquake recorded in Fennoscandia in historical times, i.e., the *c.* 5.8 magnitude in 1819 (Muir Wood 1989b, Bungum and Olesen 2005). Liquefaction structures in the postglacial overburden point to the likely occurrence of large, prehistoric earthquakes in this area. Three measurements of uplift of acorn barnacle and bladder wrack marks on the islands of Hugla and Tomma in the outer Ranafjorden area (Figure 8) show anomalously low land uplift from 1894 to 1990 (0.0–0.07m) compared to the uplift recorded to the north and south (0.23–0.30 m). Dehls et al. (2002) observed an irregular relative subsidence pattern from InSAR permanent scatterer data during the period 1992–2000 (Figure 8) in the areas with high seismicity and the observed fault scarps. The relatively low seismicity occurring at a depth of 2–12 km could therefore have created the observed irregular subsidence pattern at the surface. We have established a network of benchmarks to measure the active strain in the Ranafjorden area by use of the Global Positioning System (GPS). Three 15–20 km-long profiles were established across outer, central and inner Ranafjorden. GPS campaign measurements in 1999 and 2008 indicate that the bench marks along the western profile have moved *c.* 1 mm yr^{-1} to the NW and W relative to the stations along the two eastern profiles (Olesen et al. 2012b) (Figure 8). Fault-plane solutions indicate E–W extensional faulting (Hicks et al. 2000b).

Some of the earthquake clusters in the Handnesøya and Sjona areas are located along NNW–SSE-trending fracture zones with escarpments facing to the west. They were most likely formed by glacial plucking of the bedrock along the fractures by the moving inland ice. Ice-plucking features may, however, be indirectly related to neotectonics.

Studies of rock avalanches indicate two separate, large-magnitude earthquakes in the North Troms–Finnmark region during the period 11,000–9,000 BP (Dehls et al. 2000a, Blikra et al. 2006). There is also a possible event in the Astafjorden–Grytøya area in southern Troms where a relatively high con-

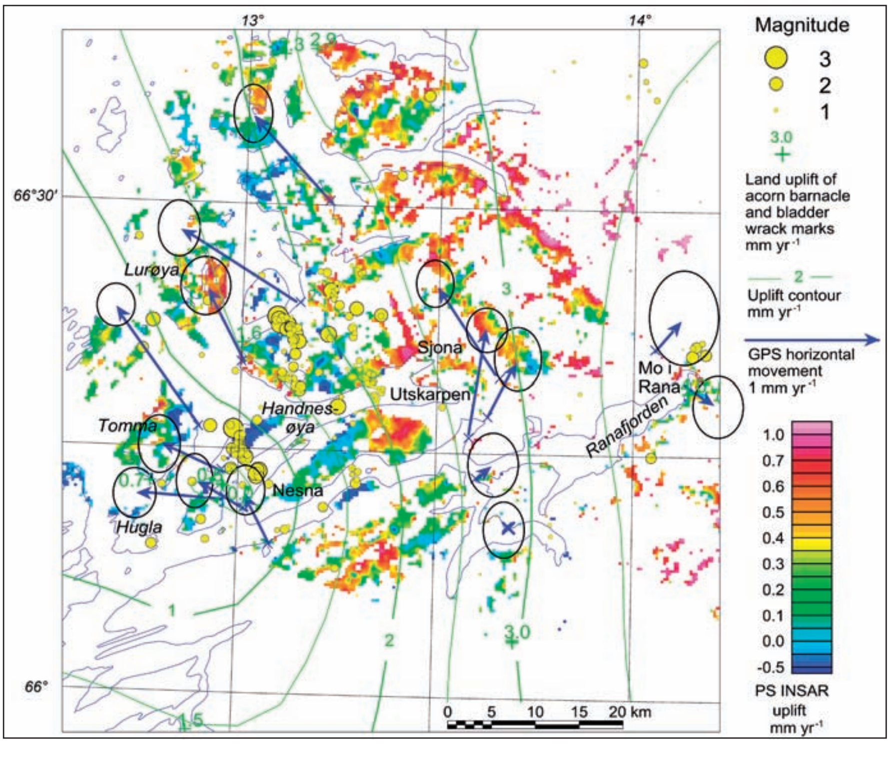


Figure 8. Annual uplift during the period 1992–2000 from the InSAR permanent scatterer method (Dehls et al. 2002). The observed seismicity from July 1997 to January 1999 (Hicks et al. 2000b) seems to occur along N–S-trending clusters that coincide with areas of relative subsidence and mapped fractures and faults with pronounced escarpments (Olesen et al. 1995). These scarps have been partly attributed to plucking effects by the moving inland ice. GPS stations to the west of the earthquake clusters have moved c. *1 mm yr^{-1} to the west relative to the stations on the eastern side during the period from 1999 to 2008 (Olesen et al. 2012b). Fault plane solutions indicate E–W extensional faulting (Hicks et al. 2000b).*

centration of rock avalanches has been recorded (Blikra et al. 2006). The latter observation has been graded as C (possibly neotectonics). Palaeoseismic events have also been postulated in western Norway (Bøe et al. 2004, Blikra et al. 2006, Longva et al. 2009). There is, for example, evidence of three regional slide events in western Norway, including one episode shortly after the deglaciation and two events at *c.* 8,000 and 2,000–2,200 calendar years BP. The 8,000 yr BP event has been attributed to the tsunami generated by the Storegga slide. An 8,000 yr BP liquefaction event registered in Nord-Trøndelag may have been triggered by an earthquake.

Offshore

Detailed analysis of offshore 2D and 3D seismic data has not yet revealed any definite neotectonic deformation structures. Several distortions in the Quaternary reflectors, however, have been mapped in the northern North Sea area. Two types of possible neotectonic features have been identified on the Norwegian continental shelf: 1) Fissures and lineaments correlated with areas of gas leakage (not obviously related to basement faults). 2) Probable reactivation of Miocene dome structures in the deeper parts of the Norwegian Sea.

The NEONOR project evaluated 14 reports of possible offshore neotectonic events. In addition, the Seabed Project assessed five neotectonic claims in the Møre and Vøring Basins (NORSAR 1999). The offshore study areas included the shelf and slope regions, but not the outermost areas overlying the oceanic crust.

Hovland (1983) described faulting of a soft, silty clay on the sea floor at the basal western slope of the Norwegian Channel. The faults terminate at shallow depths and are not connected to deeper structures. Hovland (1983) related these faults to areas of high gas saturation in the shallow sediments, and associated the structures with a release of this gas. A multibeam, echo-sounding survey (Olesen et al. 2004) carried out in 1999, within the frame of the NEONOR project, confirmed the findings of Hovland (1983). The seafloor topography in this area is characterised by N–S-trending faults and fissures with up to 1–2 m throws, and also by large, elongate pockmarks (Figure 9). Olesen et al. (2004) also reported a similar set of structures in the Kvitebjørn area located immediately to north of the bathymetric survey. In this area, there are also indications of high gas saturation at shallow depth.

Chand et al. (2012) reported a comparable set of faults in the SW Barents Sea. Unloading due to deglaciation and erosion resulted in opening of pre-existing faults and creation of new ones, facilitating fluid migration and eventual escape into the water from the subsurface. Expressions of hydrocarbon gas accumulation and fluid flow such as gas hydrates and pockmarks are widely distributed in the Barents Sea. Several gas flares, some of them 200 m high in echograms, occur along a segment of the Ringvassøy–Loppa Fault Complex, indicating open fractures and active fluid flow (Chand et al. 2012). These open fractures resemble the vertical fractures observed on mainland Nordland, which are most likely also related to Pleistocene unloading (Olesen et al. 2004, 2012b).

Faults and pockmarks similar to the ones reported from the North Sea and the Barents Sea also exist in the Storegga area on the Mid-Norwegian shelf (Fulop 1998). In these cases, it has also not been possible to relate the faults and fissures to any deeper structures. Judd and Hovland (2007) discussed the occurrence and distribution of the numerous pockmarks in relation to the present-day seismicity in the North Sea, and concluded that the seismicity was too low to have triggered a flow of fluids

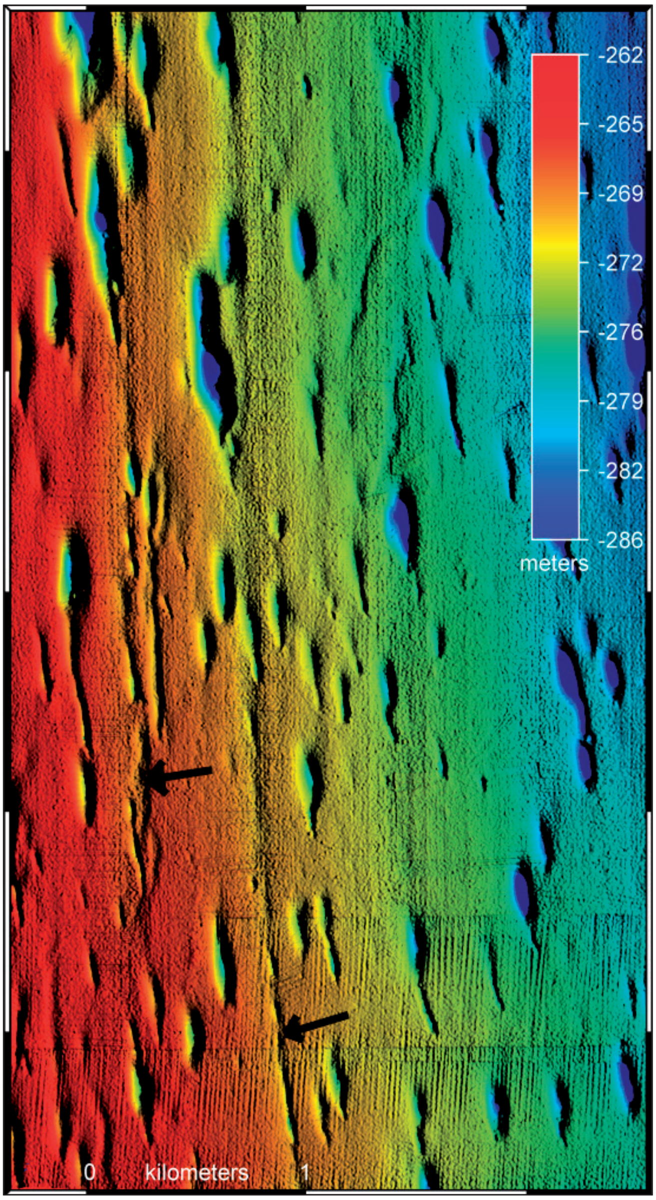


Figure 9. Bathymetry along the western margin of the Norwegian Channel south of Kvitebjørn. Abundant pockmarks (up to 500 m long and 10 m deep) occur in the area (location 28 in Figure 1 and in the Appendix). The arrows show postglacial faults, which seem to be related to the formation of the elongated pockmarks. Offset along the faults is approximately 1 m. The elongate form of the pockmarks is most likely a result of the influence of strong currents in the shallow sea immediately after the deglaciation of the area. The multibeam echo-sounding data have been acquired by the Norwegian Mapping Authority. The faults were originally reported by Hovland (1983).

and gas from the sediments. Nevertheless, one could argue that deglaciation-induced seismic pulses could have provided the necessary energy to release large quantities of gas from the North Sea sedimentary basins immediately after the last retreat of the inland ice. Bungum et al. (2005) have also suggested that large-scale postglacial earthquakes could have occurred along hidden thrusts beneath the seabed offshore Mid-Norway.

Another possible neotectonic feature that has been identified on the Mid-Norwegian continental shelf is the probable reactivation of Neogene dome structures in the deeper parts of the Norwegian Sea (Blystad et al. 1995, Vågnes et al. 1998, Lundin and Doré 2002). Contractional structures (large anticlines and synclines, reverse faults and inverted depocentres) were initiated during the Palaeogene in the Vøring and Møre Basins. There are indications that some of these structures have been growing from the Eocene to the present (Vågnes et al. 1998), with an episode of more prominent deformation in the Miocene (Lundin and Doré 2002). Doré et al. (2008) related the domal structures to the gravitational effects from the mass excess within the Iceland Plateau.

The shelf edge of the Norwegian and Barents Seas is presently a region of relatively high seismicity. Large-scale slumping also occurred along the shelf edge in the Holocene; and buried Pleistocene and older slides are common. Some slides were formed when the shelf edge was loaded by glaciers and glacial deposits, whilst others, like the main Storegga slide, are definitely postglacial. Bugge et al. (1988) and Solheim et al. (2005) speculated that earthquakes triggered the large slides. Submarine slides may, consequently, be secondary effects of neotectonic activity in some areas.

A several km-long and NNW–SSE-trending escarpment has been mapped *c.* 40 km to the SSE of Sørkapp on Svalbard (Angelo Camerlenghi, pers. comm.. 2010 on unpublished SVAIS Project multibeam data). The scarp is facing to the WSW and its height appears to be consistent. It is a candidate for a postglacial fault, but high-resolution seismic profiling is needed to validate the claim.

Seismicity

On a global scale, the seismicity of Norway is low to intermediate, even though it is the highest in northern Europe. The available historical data indicate a cumulative recurrence relation $\log(N) = 4.32–1.05M_W$ (Bungum et al. 2000), which means one earthquake of M 5 or larger every 8–9 years and one of M 6 or larger every 90–100 years. The largest earthquakes in historical times in Norway and surrounding offshore areas occurred in Storfjorden, Svalbard, in 2008, M 6.0 (Pirli et al. 2010), in the Rana region in 1819, M 5.8 (Muir Wood 1989b, Bungum and Olesen 2005), in the Vøring Basin in 1866, M 5.7, in the outer Oslofjord in 1904, M 5.4 (Bungum et al. 2009) and in the Viking Graben in 1927, M 5.3 (Bungum et al. 2003). The most recent M>5 earthquakes include an M 5.3 event in the Vøring Basin in 1988, in an area with almost no earlier seismicity (Byrkjeland et al. 2000), and an M 5.2 event in the northeastern North Sea in 1989 (Hansen et al. 1989). This indicates that we might anticipate another larger earthquake in Norway relatively soon in one of the seismically active areas, either in the Oslofjorden region or in the coastal areas of western and northern Norway, given that it is now more than 20 years since we had the last 1-in-10-year earthquake. Even so, the occurrence of earthquakes is still essentially Poisson distributed (memory free), and the location of future, large, intraplate earthquakes is also highly uncertain in a region where the causative fault is not likely to be known.

The seismicity of Norway is strictly intraplate, also along the passive continental margin, but even so it covers a region with strain rates with several orders of magnitude variation (Bungum et al. 2005, Kierulf et al. 2012) and with large variations also in tectonic conditions. The main control on the seismicity in this region may be the passive continental margin itself, with the large lateral variations in structural composition within it. Moreover, some of the large sedimentary basins (depocentres) also seem to be correlated with seismicity (especially in the Lofoten Basin), as discussed in detail by Byrkjeland et al. (2000). In the Nordland region there is also a parallel, shallow-seismic lineation along the coast, representing mostly extensional stress failure. Other seismic areas are in the failed graben structures in the North Sea and in the Oslo Rift zone (Bungum et al. 2000, Bungum et al. 2009). This pattern of seismicity is fairly consistent with the conclusions from a global study of so-called stable continental regions (SCR) (Johnston and Kanter 1990, Johnston et al. 1994, Schulte and Mooney, 2005), maintaining that rifted passive margins and failed rifts are the two main types of host structures responsible for the largest earthquakes in such areas. There are on the order of 20 such earthquakes above M 7 known to us on a global scale (Bungum et al. 2005), and recent studies from Australia (Leonard and Clark 2011, Clark et al. 2012) indicate that this number is likely to be steadily increasing. It should be kept in mind, however, that the recurrence times at any given SCR location could easily be thousands of years, in contrast to decades or centuries at plate margins. This is the situation that has given rise to recent claims that some large earthquakes in low-seismicity regions have not been 'predicted' by published hazard maps (Hanks et al. 2012). In any case, given that the largest historical earthquake on mainland Norway is on the order of M 6, it has been suggested that there may be a significant earthquake deficit in this region (Bungum et al. 2005).

Another potentially important factor for the seismicity of Norway is the fact that Fennoscandia has been fairly recently deglaciated, where we know that the initial and rapid uplift connected to this deglaciation resulted in a burst of larger earthquakes (Johnston 1987, 1989, Muir Wood, 1989a, Dehls et al. 2000a, Olesen et al. 2004), possibly even triggering the giant Storegga submarine slide (Solheim et al. 2005). We do not yet have a good understanding of the way in which the transition from the high seismicity of 10,000–6,000 years ago to the low seismicity of today has taken place, except that there are strong indications that the present-day seismicity is largely related to contemporary tectonic processes rather than being an effect of remaining glacioisostatic adjustments (Bungum et al. 2005).

Contemporary stress field

The contemporary stress field has been discussed extensively in terms of possible driving mechanisms by Fejerskov and Lindholm (2000). The discussions of potential stress-driving sources include ridge push, glacial rebound, flexural stresses through sedimentation and topography. In Norway, as well as globally, the earthquake focal mechanisms represent a unique source for understanding the underlying stresses since the earthquakes sample the deeper parts of the crust. It is, however, also important to understand the limitations, since even in regions where the global stress model is clear (e.g., in the Himalayas), each single earthquake focal mechanism may deviate significantly from the regional picture. Moreover, small earthquakes are more influenced by smaller-scale, stress-modifying factors than larger events, which carry a higher regional significance.

There is now strong evidence that the stress regime responsible for the observed seismicity appears to be the result of diverse stress-generating mechanisms at scales ranging from crustal plate to local, and that the stress field at any given place is therefore multifactorial (e.g., Bungum et al. 1991, 2005, Byrkjeland et al. 2000, Fejerskov and Lindholm 2000, Fejerskov et al. 2000, Lindholm et al. 2000, Olesen et al. 2004). Earthquakes generally occur along pre-existing zones of weakness and result from a buildup of stress and reduced effective shear strength along favourably oriented faults (Bungum et al. 2005). A key factor in reaching a better understanding of the seismicity will therefore be to improve our understanding of the interaction between the resultant stress field and the various zones of weakness in the crust.

In situ stress measurements argue for relatively high deviatoric stress magnitudes at shallow depths below the ground (Stephansson et al. 1986). The recent discovery of impressive stress-relief structures in different regions of Norway (Roberts 1991, 2000, Roberts and Myrvang 2004, Pascal et al. 2005a,b, 2006, 2010) adds support to this conclusion. Such features include reverse-fault offsets of drillholes in road-cuts and quarries, and consistently oriented, tensional axial fractures in vertical drillholes (Figure 10). Although stress deviations are observed locally in Norway, maximum principal stress axes determined both by in situ stress measurements (Figure 11) and by stress-relief features (Figure 12) are, in general, horizontal and strike NW–SE to WNW–ESE (Dehls et al. 2000b, Reinecker et al. 2005, Pascal et al. 2006), suggesting ridge-push as an important contributing mechanism (e.g., Bungum et al. 1991, Byrkjeland et al. 2000, Pascal and Gabrielsen 2001). Postglacial rebound has quite commonly been advanced as a secondary source. The viscoelastic readjustment of the lithosphere is theoretically prone to generate deviatoric stresses of a much greater magnitude than in the case of ridge-push (i.e., ~100 MPa, Stein et al. 1989). However, no clear radial pattern can be observed in the present-day stress compilations (Reinecker et al. 2005) indicating that, in contrast to the situation that prevailed just after deglaciation (Wu 1998, Steffen and Wu 2011), rebound stresses are currently relatively low. While shallow seismicity with extensional (flexural) mechanisms in the coastal regions of Nordland has earlier been associated with glacioisostatic adjustments (Hicks et al. 2000b, Bungum et al. 2005), flexure due to erosion and unloading may be a more important factor here. Flexural loading of offshore basins by high rates of sedimentation during Pliocene to Pleistocene time represents another stress source, explaining reasonably well a part of the offshore seismicity, such as in the Lofoten Basin (Byrkjeland et al. 2000), the outer part of the Mid-Norwegian shelf and the central axis of the North Sea (Figure 4). The volume of sediments deposited along the continental margin in the Pleistocene Naust Formation has been well mapped during the last decade (Figure 13, Rise et al. 2005, Dowdeswell et al. 2010) and can be used to constrain the amount and timing of onshore erosion. Average sedimentation rates during the last ice age are estimated to have been ~0.24 m kyr^{-1} with 2–3 times higher rates for the most recent 600 kyr (Eidvin et al. 2000, Dowdeswell et al. 2010). The substantial sediment erosion must have led to significant onshore exhumation and isostatic rebound. The main present-day topography, however, is considered to be much older; outcrops of deeply weathered basement rocks in the Vestfjorden and Ranafjorden areas, for example, indicate a primary inheritance from the Mesozoic (Olesen et al. 2012a).

An additional stress source that has commonly been mentioned in the literature, and tentatively quantified by Fejerskov and

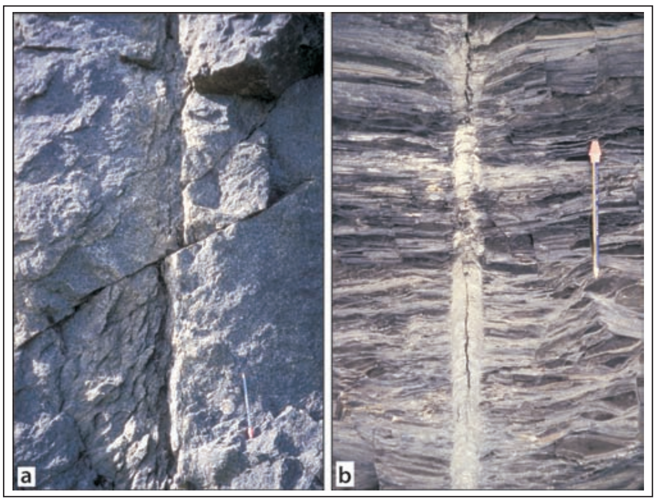


Figure 10. (a) Drillhole offset in a reverse sense along a joint surface in granulite gneisses, south of Beskelandsfjorden, Roan, Fosen Peninsula, Sør-Trøndelag; looking south. Locality – 1:50,000 map-sheet 'Roan' 1623 III, 3–NOR edition, grid-ref. NS6075 1815. (b) Well developed axial fracture in a drillhole in slates of the Friarfjord Formation, Laksefjord Nappe Complex, from the roofing slate quarries at Friarfjord, close to the old quay. 1:50,000 map-sheet 'Adamsfjord' 2135 I, 3–NOR edition, grid-ref. MU9695 1810. This particular quarry face trends N–S, and the photo is taken looking due west.

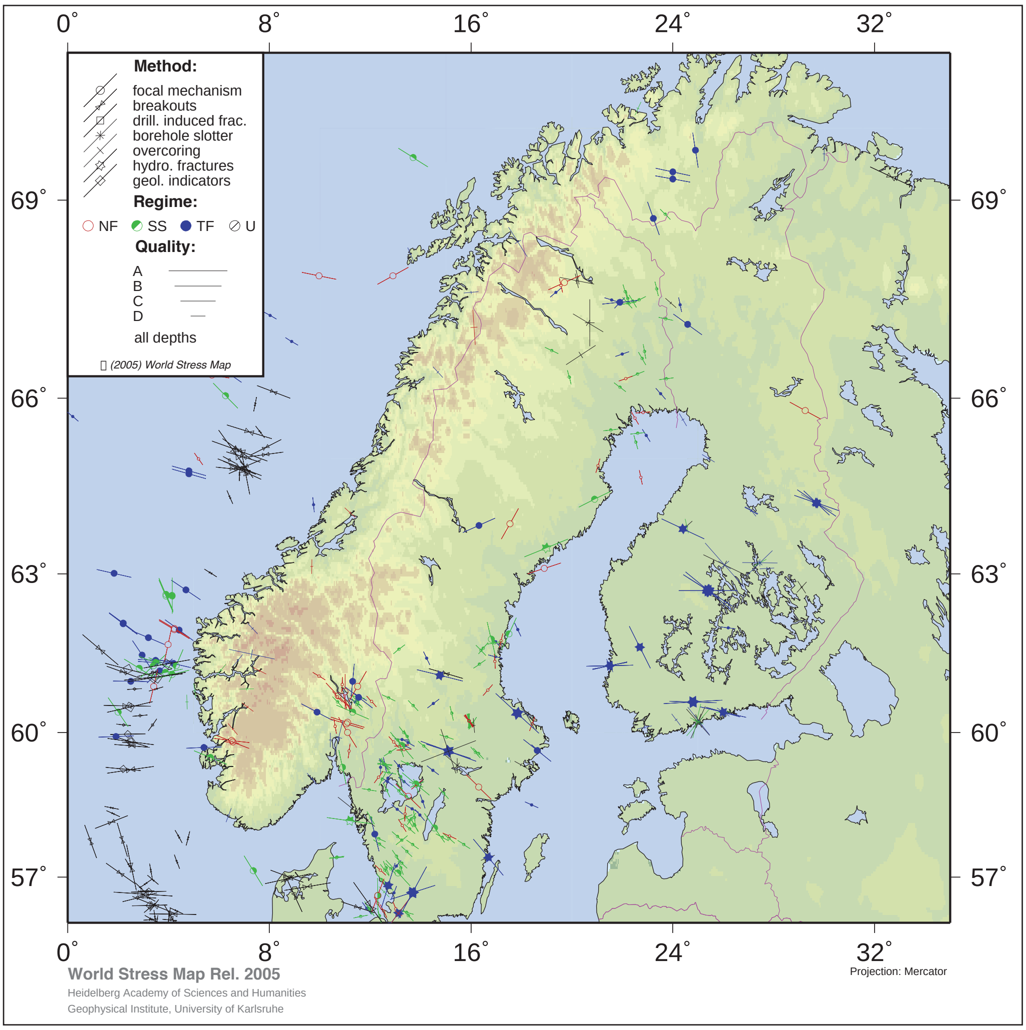


Figure 11. Contemporary stress orientations in Fennoscandia taken from the World Stress Map (Reinecker et al. 2005). Note that S_{Hmax} is predominantly NW–SE oriented.

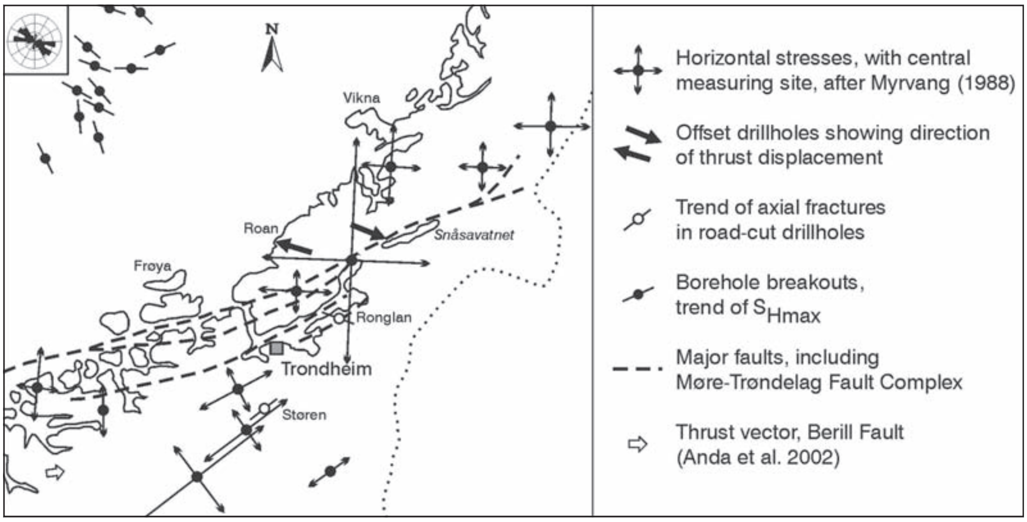


Figure 12. Outline map showing the diverse rock-stress orientation data from central Norway and the Trøndelag Platform. The small rose diagram (inset, top left) is from Hicks et al. (2000a, p. 243) and depicts the trends of maximum horizontal compressive stress as derived from earthquake focal mechanism solutions in the area of offshore Mid Norway (period 1980–1999). The figure is from Roberts and Myrvang (2004).

Lindholm (2000), is associated with the anomalous elevation differences of southern and, to a lesser extent, northern Norway. It has been shown recently, for example, that the southern Scandinavian mountains are likely to generate significant gravitational stresses in adjacent offshore sedimentary basins (Pascal and Cloetingh 2009). This model offers an alternative explanation to the anticlockwise stress rotation observed from the Norwegian margin to the northern North Sea, which Fejerskov and Lindholm (2000) found to be consistent with gradually changing ridge-push directions. The NE–SW stress orientations detected southeast of the Møre–Trøndelag Fault Complex (Figure 12) (Roberts and Myrvang 2004) have also been interpreted (Pascal and Cloetingh 2009) in terms of changes in gravitational stresses. In the Oslo Region and Nordland (Ranafjord area, see Hicks et al. 2000b), the stress patterns appear to be more complex, probably simply because there are more observations from these regions. In the Oslo Region, the orientation of the maximum horizontal stress axis is, in general, WNW–ESE, but with local deviations and stress permutations (Hicks et al. 2000a, Dehls et al. 2000b, Pascal et al. 2006, 2010). There is, for example, a (weak) tendency for focal mechanisms of shallow (<13 km) earthquakes to relate mostly to normal faulting, whereas deeper events indicate strike-slip and reverse faulting (Hicks 1996). It is tempting to interpret this complex stress pattern in terms of flexuring due to Neogene erosion and unloading and, perhaps, in terms of structural complexity (including lateral changes in rheology) inherited from the Permian magmatic and rifting event.

In Nordland, as mentioned earlier, inversion of focal mechanisms of earthquakes indicates a coast-perpendicular extensional stress regime with shallow earthquakes (Figure 14), which is directly opposite to what is found along the margin farther offshore (Hicks et al. 2000b, Bungum et al. 2010). There are, however, also some strike-slip earthquakes here, with coast-parallel compressions. This anomalous stress field (contrasting with the regional one) appears to be associated with a locally enhanced uplift pattern and a related flexuring mechanism. This may in turn be related to remaining glacioisostatic adjustments, but since very recent erosion has taken place in Nordland, the crust there may be strongly flexed, which also would result in coast-perpendicular extension.

In Trøndelag, central Norway, rock-stress measurements and stress-release features have shown that the Møre–Trøndelag Fault Complex marks an important structural divide separating crustal blocks with disparate, present-day stress fields (Roberts and Myrvang 2004), as previously suggested by numerical-modelling studies (Pascal and Gabrielsen 2001). A NW–SE to WNW–ESE horizontal compression prevails in coastal areas northwest of the fault complex, and accords with borehole breakout and earthquake focal mechanism solution data acquired offshore (Figure 11). This linked stress pattern, from onshore to offshore, provides further support for the notion that the dominant S_{Hmax} trend is likely to relate to a distributed ridge-push force arising from divergent spreading along the axial ridge between the Norwegian and Greenland Seas.

Some important points, however, should be kept in mind here. Firstly, small earthquakes have a similarly small regional

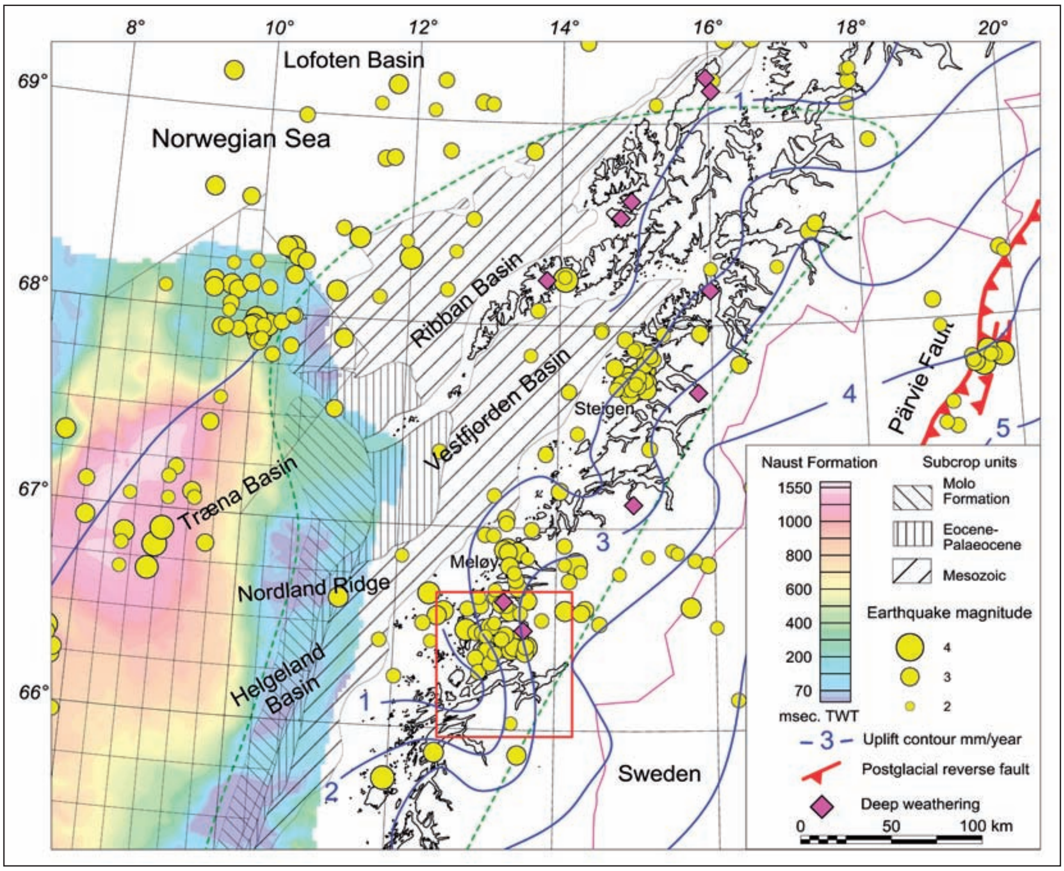


Figure 13. Map of the Nordland margin, including source catchment area of glacial erosion (green dashed line) and area of offshore deposition (isopach map of thickness of the Naust Formation in milliseconds of two-way travel time, where 1 ms is ~1 m). Blue line marks present shelf edge. Adapted from Dowdeswell et al. (2010). Zones of deep weathering up to 100 m thick and more than 10 km wide occur within the eroded area indicating that the present landscape is to a large degree of Mesozoic age. Subcrop units (modified from Sigmond 2002) underlying the Naust Formation are mainly Tertiary, Cretaceous, and Jurassic sedimentary rocks (hatched pattern). Earthquakes from the period 1989–2011 are shown in yellow and the size of the circles reflects the magnitude on the Richter scale. The red frame depicts the location of Figure 8.

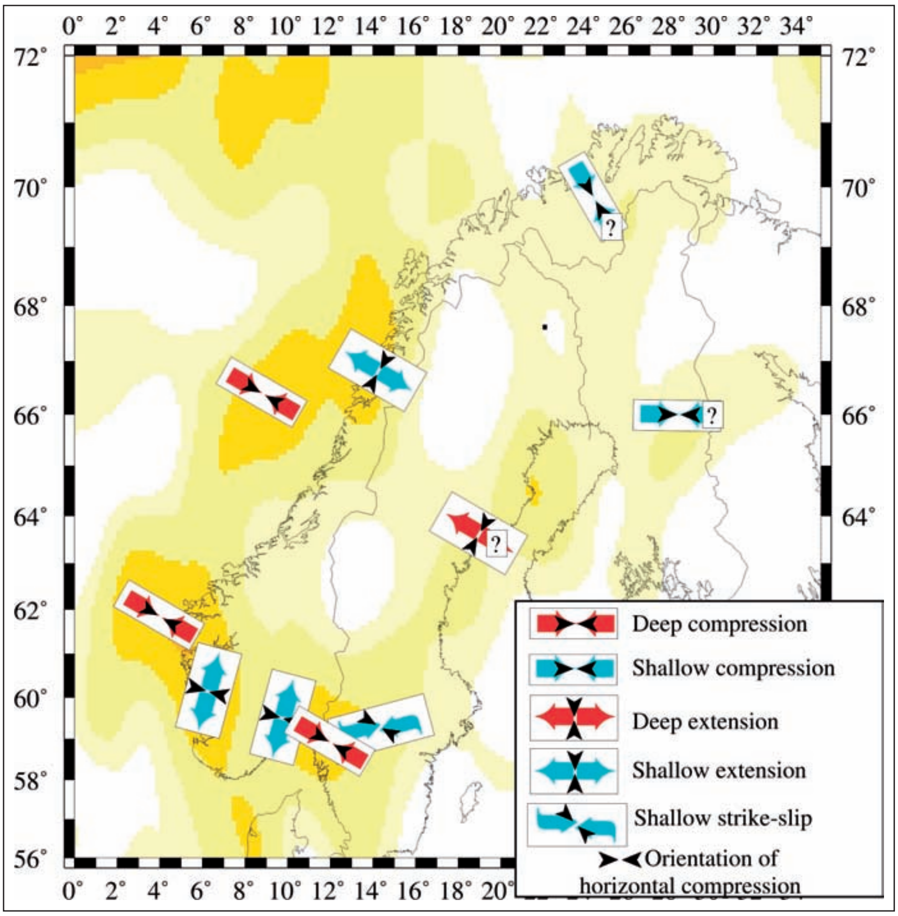


Figure 14. Stress orientations, type of faulting and focal depths synthesised from earthquake focal mechanisms and in situ stress measurements (from Fjeldskaar et al. 2000). Areas with sparse data are indicated with question marks. Intensity of yellow indicates intensity of seismicity. Note that offshore depocentres generally coincide with areas of dominating compressional events whereas the coastal areas have a predominantly extensional regime.

significance and their inferred stress orientations may deviate from the regional stress pattern for a variety of reasons, also because of the assumption that the stress axes bisect the angles between the nodal planes and because earthquakes in general occur along pre-existing faults (e.g., Sibson 1990). Secondly, the crustal stress tensor everywhere is built up with contributions from a number of sources and therefore cannot be explained by a single contributing source. It is, therefore, surprising that the stress orientations are as stable and consistent as they have been shown to be.

Postglacial and contemporary uplift

A dataset of the absolute vertical uplift of Fennoscandia compiled from tide gauges, precise levelling and continuous GPS stations (Vestøl 2006) has been combined with seismicity recorded during the period 1980–2011 and is shown in Figure 15a. The figure shows no clear correlation between onshore uplift and seismicity in Fennoscandia. However, in a model that combines offshore subsidence with onshore uplift, it is readily understood that the coastal regions will be relatively most susceptible to crustal flexuring and deformation, as also confirmed by present-day seismicity. The BIFROST GPS network (Milne et al. 2001, Lidberg et al. 2007) offers a regional 3D image of the bedrock deformation within the Fennoscandian Shield and provides, for the first time, information on the horizontal movement of the bedrock. Both datasets show that the first-order deformation is dominated by the glacial isostatic adjustment. The maximum vertical uplift of 11.2 ± 0.2 mm yr^{-1} occurs in the Umeå area (Milne et al. 2001, Scherneck et al. 2001, 2003, Lidberg et al. 2007). The horizontal movements are directed outward from this location on all sides with the highest values located to the northwest and east (reaching 2 mm yr^{-1}). The northwestern area coincides with the Lapland province of postglacial faulting in northern Fennoscandia. Kakkuri (1997) also measured a maximum, present-day, horizontal strain in the region of postglacial faults in northern Finland. Pan et al. (2001), however, reported differential horizontal displacements along the border zone between the Fennoscandian Shield and the European lowland.

Semiregional deviations from the regional uplift pattern in the order of 1–2 mm yr^{-1} have been reported by Olesen et al. (1995) and Dehls et al. (2002) for the Ranafjorden area in northern Norway. This conclusion is deduced from two independent datasets, namely repeated levelling and permanent scatterer techniques. Fault-plane solutions reported by Hicks et al. (2000b) show extensional faulting in the same area. Vestøl (2006) has carried out a least-squares collocation adjustment of the combined precise levelling, tide gauge recordings and time series from continuous GPS stations. He concluded that some semiregional uplift anomalies in Fennoscandia are related to inaccuracies in the original levelling data.

The Fennoscandian Shield was affected by a Neogene phase of passive doming (approximately 1,000 m amplitude) in southern Norway and in the Lofoten–Troms area (Riis 1996). Hence, the present elevation of Scandinavia is partly the result of Neogene uplift and exhumation of a fault-controlled topography (Osmundsen et al. 2010). The combined effect of tectonic uplift of Fennoscandia and the onset of the northern hemisphere glaciation led to greatly increased erosion and sedimentation. More than 50% of the volume of Cenozoic sediments was actually deposited during the last 2.6 m.y (last 5% of the time period).

There is some evidence (e.g., Mangerud et al. 1981, Sejrup 1987) that the Norwegian coast may have been subject to tectonic uplift of the order of 0.1–0.3 mm yr^{-1} during the Quaternary, in addition to postglacial uplift, as also suggested by Mörner (1980). Recent studies of uplifted, Middle and Upper Weichselian, marine sediments (Olsen and Grøsfjeld 1999) show, however, that the inland ice sheet fluctuated quite frequently during the 50,000–18,000 yr BP interval. Repeated and rapid ice retreat following heavy ice loading was the most likely mechanism for depositing marine sediments of both the same and different age intervals in several uplifted positions along the coast of Norway as well as in inland areas of southeastern Norway. This process can also explain the presence of elevated Weichselian marine clay on Høg–Jæren and coastal caves above the maximum Holocene marine limit on the innermost strandflat in western and northern Norway. These elevated caves have also been interpreted in terms of a Neogene tectonic uplift (Holtedahl 1984, Sjöberg 1988).

Fjeldskaar et al. (2000) argued that the long-term Neogene uplift of western Scandinavia is still active and can explain approximately 1 mm yr^{-1} of the present uplift of the southern and northern Scandinavian mountains. Geodynamic modelling of the present and postglacial uplift data shows that the bulk of the present-day uplift can be explained as a response to glacial unloading (Fjeldskaar et al. 2000). The model for uplift within three specific areas deviates, however, from the observed uplift: 1) a zone including northwestern Norway and part of eastern Norway, 2) the Lofoten–Troms area, and 3) the Bay of Bothnia area. The Bothnia area shows a negative deviation between the observed and calculated uplift whereas the two Norwegian

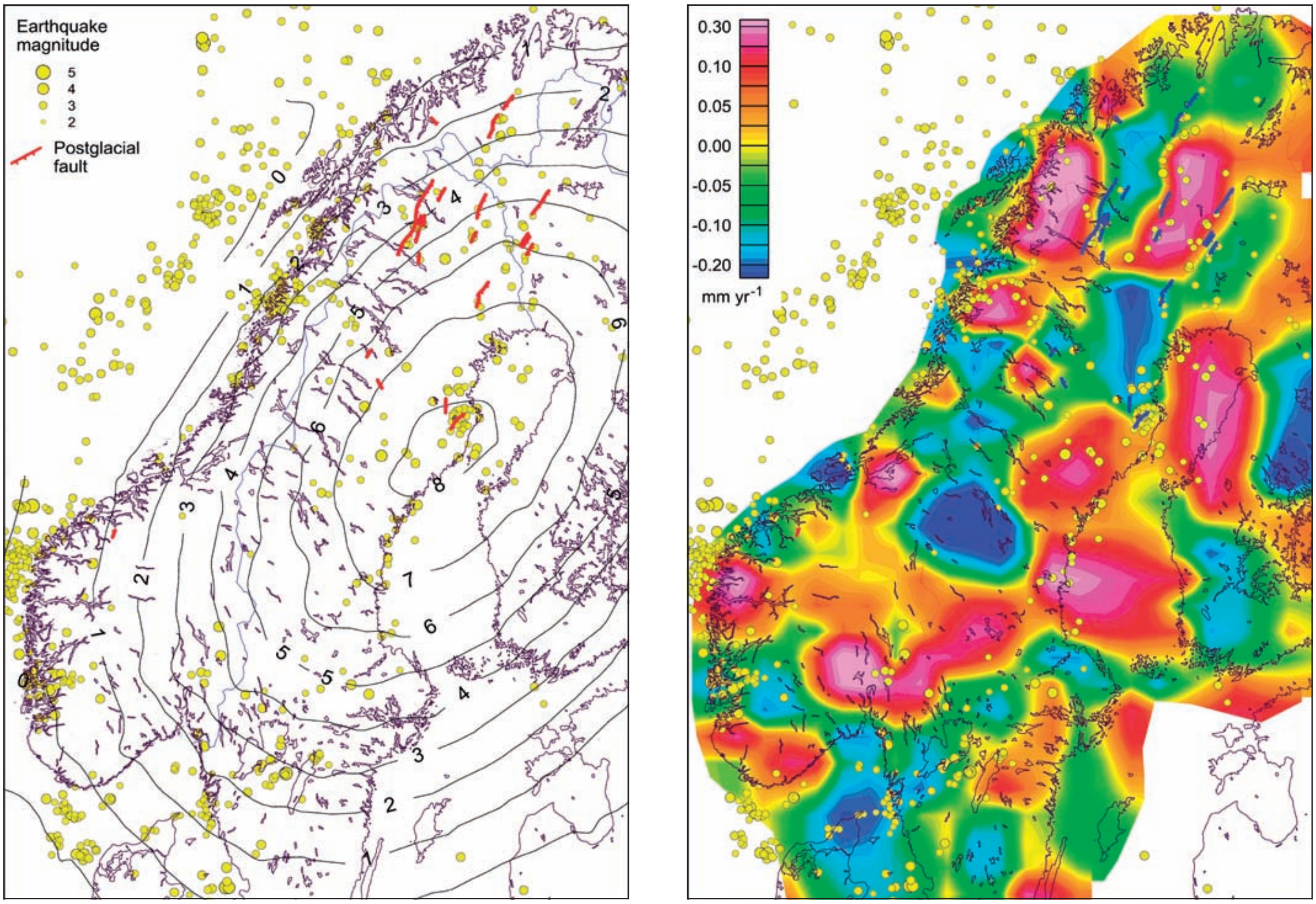


Figure 15. Present-day annual velocity of the Fennoscandian bedrock. (a) Amount of uplift in mm per year (Vestøl 2006) and 1980–2011 earthquake epicentres in Fennoscandia. There is no direct correlation between uplift pattern and seismicity in Fennoscandia (Bungum et al. 2010). (b) Deviation from a fifth-order polynomial trend surface of the present-day annual uplift in A) (Vestøl 2006). The anomalies in the order of ±0.3 mm yr^{-1} could represent systematic or random noise, tectonic components or deviations in the uplift pattern as a result of thickness variations of the inland ice (Bungum et al. 2010).

areas show positive deviations. The two areas in Norway also coincide partly with the Neogene domes in southern Norway and Lofoten–Troms, indicating that a long-term tectonic component is partly responsible for the present-day uplift.

Vestøl (2006) compiled uplift data from precise levelling, tide gauge recordings and time series from continuous GPS stations applying the Kriging adjustment method, and concluded that it is, in general, hard to claim uplift anomalies with wavelengths shorter than 50–100 km and amplitudes greater than ±0.4 mm in Norway (Figure 15b). Vestøl (2006) also subtracted the modelled glacial isostatic uplift determined by Lambeck et al. (1998) from his own postglacial uplift data and arrived at a similar conclusion as Fjeldskaar et al. (2000). The three anomalous uplift areas are, however, larger and reveal a larger wavelength component than the anomalous areas of Fjeldskaar et al. (2000). The anomalous uplift of about 0.4 mm yr^{-1} over a *c.* 50 km-wide area in the Oslo region also seems to be significant (Vestøl 2006). There is, however, a need to check the reliability of the semiregional uplift anomalies produced with the InSAR permanent scatterer technique (Bungum et al. 2010). Uplift anomalies of several millimetres in the Lyngen area, Troms (Osmundsen et al. 2009), are, for instance, inconsistant with the anomalous uplift dataset produced from repeated levelling). The recorded seismic events in outer Lyngenfjorden are most likely related to blasting during the molo construction at Årviksand harbour in the winter of 1999. The recorded negative arrivals from these events support this alterative interpretation. The small-magnitude events in the Norwegian earthquake catalogue are contaminated by explosions.

There have been earlier reports (from geodetic measurements) of contemporary movements along a fault in Ølen in southwestern Norway (Anundsen 1989) and along the Stuoragurra Fault in Finnmark (Olesen et al. 1992a). Repeated levelling within the NEONOR project failed, however, to provide any support for aseismic movements along either of these faults (Sylvester 1999, Bockmann and Grimstveit 2000). It is, therefore, suggested that most of the local anomalies in the uplift pattern are related to inaccuracies in the levelling data (Olesen et al. 2004). The intermediate component of the regional uplift pattern (anomaly areas with *c.* 100 km wavelength and 0.5–1 mm yr^{-1}) may, however, be related to tectonic processes other than glacioisostasy.

Discussion

A number of Late Quaternary to Holocene deformation features in Norway can be associated with earthquakes. Sackung structures, for example, occur as a result of bidirectional extension of the Kvasshaugen ridge in Beiarn, Nordland. Clefts up to 20 m wide and 10 m deep can be followed along an approximately 5 km-long, NNE–SSW-trending zone (Grønlie 1939, Olesen et al. 2004). The reverse Berill Fault at the foot of the Middagstinden mountain is associated with open clefts with upward-facing scarps along the mountain ridge in the hanging-wall block, which are also features typical of gravity-induced sackung structures.

Similar sackung structures, consisting of double-crested ridges, linear troughs and both upslope-facing and downslope-facing scarps, occur in the Alps, the Rocky Mountains and in New Zealand (Beck 1968, Zischinsky 1969, Savage and Varnes 1987, Varnes et al. 1989). These characteristic geomorphological forms are considered to be produced by gravitational spreading of steep-sided ridges (Varnes et al. 1989). Ambrosi and Crosta (2006) conclude that sacking structures are complex mass movements controlled by many different factors: recent geological history (e.g. glaciation and deglaciation, periglacial conditions, erosion and valley deepening, tectonic stress, uplift, seismicity, and landsliding); structural features (joints, faults, foliations); slope materials (lithology, weathering, and metamorphism); topographic factors (slope length and gradient); and groundwater conditions. Whether initiation of the movements is by strong ground-shaking (earthquakes), faulting, long-term creep, or a combination of these factors, has long been a matter of debate (Jibson 1996, Ambrosi and Crosta 2006). Varnes et al. (1989) and McCalpin and Irvine (1995) argued that the movement originates from long-term, gravity-driven creep, but the former authors exclude earthquake shaking as a possible contribution. Other investigations in New Zealand, Slovakia, Russia and Italy have concluded that earthquake shaking was a likely fault-triggering mechanism (Jahn 1964, Beck 1968, Jibson 1996, Ambrosi and Crosta 2006), partly because the sackung features occur in seismically active areas in these particular cases. Kvasshaugen is also situated in a seismically active area; hence, earthquake shaking could have been the triggering mechanism (Olesen et al. 2004). Similar structures on Øtrefjellet in Haram, Møre & Romsdal, are located on the margins of a seismically active area (Anda et al. 2002, Braathen et al. 2004, Blikra et al. 2006). Oppikofer et al. (2008) have suggested that sections of the Åknes rockslide in western Norway represent sackung structures. Ambrosi and Crosta (2006) argue that such slow, deep-seated gravitational movements can damage or destroy infrastructure, and sections of individual sackung may accelerate during rainstorms or with climate change.

In southern Norway, there is no conclusive evidence for any postglacial faulting, even though the majority of the original neotectonic claims were reported from this region. There are, however, other indirect palaeoseismic indications along a NNE–SSW-trending zone from Odda in Hardanger to Aurland in Sogn and on the coast of Møre & Romsdal, seen as a series of rock-slope failures, many in relatively gently dipping terrain (Braathen et al. 2004, Blikra et al. 2006). The former zone is situated within the Mandal–Molde lineament zone of Gabrielsen et al. (2002). Liquefaction and semiliquefaction structures in sand located close to the intersection of two topographical lineaments in Flatanger, Nord-Trøndelag (site 11 in Figure 1 and Appendix), have also been taken as evidence of a large postglacial earthquake (Olsen and Sveian 1994). There is a general lack of Quaternary overburden in western Norway and, consequently, there are few means of identifying and dating movements along the abundant bedrock scarps in the area. We cannot, therefore, totally rule out the possibility of more extensive postglacial faulting in this region.

It is considered likely that the postglacial faults, and the varying stress fields and vertical uplifts, were caused mainly by an interaction of the ridge-push force from the Knipovich and Mohn ridges and other processes such as postglacial rebound, gravitational effects from mass excess within the Seiland Igneous Province, the Lofoten–Vesterålen metamorphic complex and mountainous areas in Scandinavia (Muir Wood 1989a, 2000, Olesen 1991, Olesen et al. 1992a, Bungum and Lindholm 1997) and sedimentary loading (Byrkjeland et al. 2000, Hicks et al. 2000b) and unloading in the coastal and offshore areas. Considering that the Pliocene/Pleistocene loading of the relatively stiff oceanic crust causes seismicity in the Norwegian Sea, it is also likely that a comparable unloading of the coastal areas in western and northern Norway may induce earthquake activity. It has also been argued by Lidmar-Bergström (1995) and Lidmar-Bergström et al. (1999) that the coastal areas of southeastern Norway, together with the Skagerrak and Kattegat, were exhumed in the Neogene. Unloading of the crust in these areas can therefore also partly explain the observed seismicity in that region.

In this context it is also important, however, to appreciate the potential significance of aseismic movements which, in fact, are found over a wide range of strain-rate conditions. Recent research in subduction zones has demonstrated the importance of so-called silent earthquakes, filling the gap between brittle and ductile deformation. One effect of this is to generate significant variations in coupling ratios, relating seismic moment rates to geological moment rates. For example, following the 2004 M 9.2–9.3 Sumatra earthquake, more than 20 cm of afterslip has been documented in that region. Even though rates and scales are greatly different in Fennoscandia, there are reasons to believe that some of the same principles may apply with respect to the range of deformational processes involved.

A related question here is whether earthquakes and earthquake swarms, which in Norway are usually quite shallow, should be considered as neotectonic triggering phenomena in cases where they are not specifically related to mapped faults. This is really a question of definition, since such activities clearly also reflect movements on faults, albeit at different scales in time and displacement than are otherwise considered as neotectonics.

Isostatic modelling by Olesen et al. (2002) and Ebbing and Olesen (2005) indicates that the mountains in southern Norway are supported by low-density rocks at typical Moho and sub-Moho depths. The gravity field in the northern Scandinavian mountains, on the other hand, seems to be compensated by low-density masses at a relatively shallow depth in the upper crust. The results are in agreement with the conclusions of Riis (1996) and Lidmar-Bergström et al. (1999) that the southern Norwegian plateau was partly uplifted in the Neogene, whereas the northern mountains originated mainly as a rift-shoulder in Late Cretaceous to Early Tertiary times. In this regard, Hendriks and Andriessen (2002) reported that apatite fission-track analytical data along a profile from Lofoten into northern Sweden fit best with those expected from a retreating scarp model.

The distinct concentration of gravitational faults and slope failures in the Odda–Aurland area (site 17 in Figure 1 and Appendix) in Sogn & Fjordane and Hordaland, and in parts of Møre & Romsdal and Troms, may indicate the occurrence of large-magnitude prehistoric earthquakes in these areas (see discussions in Braathen et al. 2004 and Blikra et al. 2006). An abundance of liquefaction structures in the postglacial overburden in the Ranafjord area and the clusters of rock-slope failures in western Norway also point to the likely occurrence of large, prehistoric earthquakes in these areas. One conclusion that can be drawn from these observations and inferences is that palaeoseismology will be an important field of further research in Norway. It will be critical to date the individual rock avalanches, landslides and liquefaction events to determine whether or not they can be related to large-magnitude earthquakes.

Morsing (1757) concluded that the Tjellefonna and Silset rock avalanches on 23 February 1756 in Møre & Romsdal county were caused by an earthquake (Bungum and Lindholm 2007, Redfield and Osmundsen 2009). The Tjellefonna rock avalanche created a 40 m-high tsunami that caused catastrophic damage to the settlements along Langfjorden and Eresfjorden. A total of 32 people perished in the disaster (Schøning 1778). A landslide and several smaller rock avalanches were triggered by the 1819 M 5.8 earthquake in the Rana–Sjona area (Heltzen 1834, Muir Wood 1989b, Bungum and Olesen 2005). Although no lives were lost in this disaster, the landslide and one of the rock avalanches destroyed substantial areas of farmland (Heltzen 1834, Muir Wood 1989b, Bungum and Olesen 2005). We therefore conclude that rock avalanches and landslides, potentially triggered by future earthquakes, would almost certainly generate tsunamis in fjords and lakes and consequently constitute the greatest seismic hazard to society in Norway. The giant 1958 landslide at the head of Lituya Bay in Alaska was triggered by an earthquake and generated a wave with an initial amplitude of 524 m (Miller 1960). The mega-tsunami surged over the headland opposite, stripping trees and soil down to bedrock, and surged along the fjord which forms Lituya Bay, destroying a fishing boat anchored there and killing two people.

Trenching across the Stuoragurra Fault has shown that fault breccia was injected over a horizontal distance of more than 12–14 m from the fault zone into the lower part of the glacial overburden (Figure 6 and Dehls et al. 2000a). In order for such an injection to occur, the fault breccia must have been fluidised with high-pressure groundwater or gas. This observation shows that a sudden flow of fluids or gas can locally be associated with seismic pulses. Muir Wood (1989a) emphasised the role of pore-water pressure for reducing the effective stress in the upper part of the crust during deglaciation. Lagerbäck and

Sundh (2008) pointed out the possible significance of an assumed overpressuring of pore water beneath a thick layer of permafrost in the formation of the young faults. These same authors have also argued that the modest glacial erosion during the Weichselian would have allowed fault scarps of a similar magnitude to the Fennoscandian postglacial faults to be preserved, had they occurred. Following the mechanism that large continental ice sheets suppress the release of earthquakes during glaciations (Johnston 1987), it is still likely that previous deglaciations were accompanied by large earthquakes though not necessarily the size of those in post-Weichselian time.

Erosion and uplift of offshore areas along the coast of Norway and in the Barents Sea would most likely release gas from hydrates within the sediments (Chand et al. 2012). The climatic implications of the deglaciation-related seismicity is therefore of special interest. Seismic pumping (Sibson et al. 1975) and the release of hydrocarbons during deglaciation-induced seismic pulses may partly explain the improved climate immediately after the Weichselian deglaciation as a result of an increased greenhouse effect (Olesen et al. 2004). The greenhouse effect of CH_4 is 23 times greater than that of CO_2 (Forster et al. 2007). Paull et al. (1991) presented a scenario of a *c.* 100 m sea-level fall associated with the Pleistocene glaciations, leading to gas hydrate instability and major slumping on the continental margins. The release of large quantities of methane into the atmosphere could have eventually triggered a negative feedback to advancing glaciation once the methane emissions increased over a threshold level, leading to termination of the glacial cycle. This process could, moreover, explain why glaciations generally terminate rather abruptly.

We would also like to stress that although the level of seismicity in Norway is stable, the physical and societal vulnerability to earthquakes has increased enormously over historical time, like in most other parts of the world. When comparing the population and infrastructure of the Rana region of the early 1800s, where the 1819 M 5.8 earthquake (Muir Wood 1989b, Bungum and Olesen 2005) occurred, with the population and infrastructure of the same area today, we realise how many more people and elements of modern infrastructure are at risk. The population has increased from 3,000 to 30,000. Very few roads, no industry and mostly one-storey wooden houses existed in the Rana area in 1819 whereas today we find a major road system, railways, hydropower plants, bridges, tunnels, smelters and tall buildings. The 1992 M 5.8–5.9 earthquake at Roermond, in the Netherlands, for example, cost Dutch society about 100 million Euros (Berz 1994, van Eck and Davenport 1994). Even though the geology and population density in this industrialised flood plain area is not directly comparable to Norway, the Roermond earthquake still illustrates the importance of carrying out state-of-the-art seismic-hazard analyses also in Norway.

Kukkonen et al. (2010) have suggested that we should explore the postglacial faults in northern Fennoscandia through scientific drilling and study their characteristics, including structure and rock properties, present and past seismic activity and state of stress, as well as hydrogeology and the associated deep biosphere. This cumulative research is anticipated to advance our knowledge of neotectonics, hydrogeology and the deep biosphere, and provide important information for nuclear waste disposal, petroleum exploration on the Norwegian continental shelf and safety of hydropower dams and other infrastructure.

Conclusions

While firm evidence for postglacial deformation has been found in northern Norway, we have not yet been able to find similar evidence in southern Norway. Indications of neotectonic deformation include, however, secondary effects of possible large prehistoric earthquakes such as liquefaction and semiliquefaction structures in the Flatanger (Nord-Trøndelag) and Rana (Nordland) areas, and gravitational spreading and faulting features (sackung) on Kvasshaugen in Beiarn (Nordland) and at Berill in Rauma and Øtrefjellet in Haram (Møre & Romsdal). A series of gravitational fault systems and large rock avalanches in zones from Odda to Aurland (Hordaland and Sogn & Fjordane) and in northern Troms are also included in the grade B group (probably neotectonics). Gravitational processes primarily control the large-scale rock avalanches, mountain-ridge spreading and normal faulting, but their spatial occurrence and locations on relatively gentle slopes indicate that other mechanisms were also involved. Extra loading due to strong ground-shaking from large-magnitude earthquakes might have been an important factor.

The observed neotectonic deformation in Norway supports previous conclusions regarding a major seismic 'pulse' (with several magnitude 7–8 earthquakes) which followed immediately after the Weichselian deglaciation of northern Fennoscandia (Lagerbäck 1990, 1992, Kuivamäki et al. 1998, Lagerbäck and Sundh 2008). The 80 km-long Stuoragurra Fault constitutes the Norwegian part of the Lapland Fault Province, which consists of ten NE–SW-striking reverse faults (Table 1). Trenching of the Stuoragurra Fault in Masi has revealed that most of the 7 m-high scarp was formed in one seismic event (M 7.4–7.7) during the very last part of the last deglaciation in Finnmark (i.e., *c.* 9,300 years BP) or shortly afterwards (Dehls et al. 2000a).

Large continental ice sheets suppress the release of earthquakes (Johnston 1987). There were, therefore, most likely major seismic pulses in mainland Fennoscandia accompanying each of the deglaciations that followed the multiple glaciation cycles during the last 600,000 years. It is possible that seismic pumping (Sibson et al. 1975) associated with these glacial cycles may have assisted in extracting hydrocarbons from their source rocks and pumping them to reservoir formations and further, through gas chimneys, to produce pockmarks on the sea floor (Muir Wood and King 1993, Olesen et al. 2004).

There is some evidence from uplifted pre-Weichselian sediments and caves along the coast of western Norway that the

Norwegian coast may have been subject to tectonic uplift of the order of 0.1–0.3 mm yr^{-1} during the Quaternary, in addition to postglacial uplift. Unloading of the coastal area of Norway and loading of the outer shelf and deep water areas in the Norwegian Sea are a likely contributor to this tectonic uplift. A significant part of the seismicity along the Norwegian continental margin occurs within these two, distinctly different, tectonic regimes.

The severe Pliocene/Pleistocene uplift and erosion of coastal areas of Norway and the Barents Sea have effects on the petroleum exploration. Where uplift and tilting resulted in local extension, seal breaching and spillage may also have occurred, as in the Barents Sea (Nyland et al. 1992). The cooling of the source rocks owing to vertical movement may also cause hydrocarbon generation to decrease.

We conclude that the distribution of earthquakes, in situ rock stress, Neogene and present uplift and postglacial faulting in Norway and along the Norwegian margin seem to be primarily related to gravitational effects of excess mass along the Mohns Ridge, within the Iceland Plateau and in the southern Scandinavian mountains, to Pliocene/Pleistocene sedimentary loading/unloading, and also to postglacial rebound.

We also conclude that magnitude 6+ earthquakes are possible today in the most seismically active areas in Norway, such as the coastal parts of western Norway, Nordland and the Oslo rift zone. Rock avalanches and landslides, potentially triggered by earthquakes, could generate tsunamis and thus constitute a significant seismic hazard to society in Norway.

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Appendix

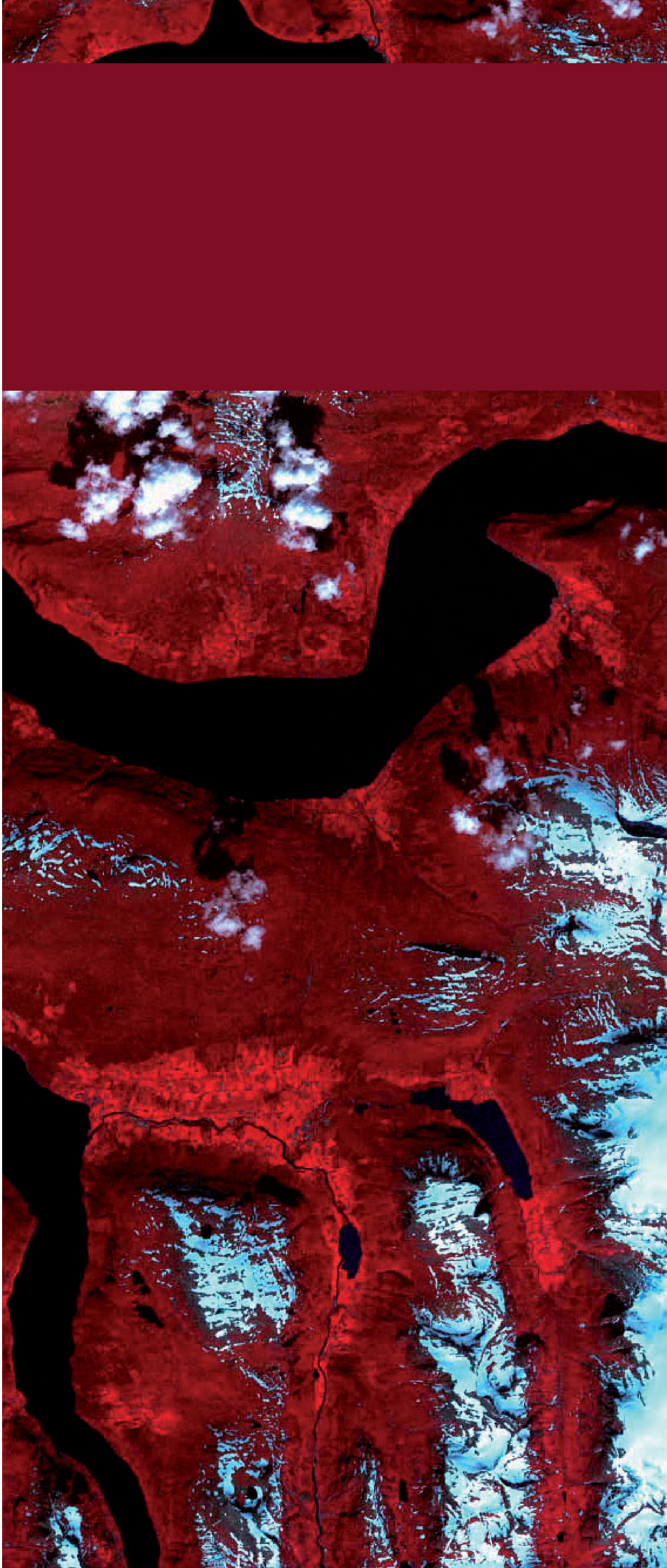
Reported evidence of neotectonics in Norway and Svalbard and assessments of the claims. The locations are ordered from north to south and are shown in Figure 1. The criteria for classification of postglacial faulting proposed by Fenton (1994) and Muir Wood (1993) have been adapted for grading the claims into the classes: (A) Almost certainly neotectonics, (B) Probably neotectonics and (C) Possibly neotectonics. Grade D, Probably not neotectonics and E, Very unlikely to be neotectonics, are omitted in the present table but are listed in Olesen et al. (2004). The most likely nature of the proposed neotectonic deformation has been included as 'TYPE' in the fifth column: (1) Tectonic fault, (2) Gravity-induced fault, (3) Erosional phenomena, (4) Overburden draping of bedrock features, (5) Differential compaction, (6) Stress release features, (7) Inconsistent shoreline correlation, (8) Unstable benchmarks or levelling errors, (9) Insufficient data resolution.

Loc.No.	Location and reference	Observation	Comment	Grade/ type
1	Bockfjord, Lihøgda, Svalbard (Piepjohn 1994)	A N–S-trending, c. 2 km-long escarpment in the Devonian sediments on the western shore of Bockfjord (an arm of Woodfjord). The apparent downthrow is to the east.	The fault scarp is linear and has a rather consistent height. The overburden, however, is thin and consists mostly of weathered bedrock. Dating of the scarp is therefore difficult.	C1, 3
2	Øksfjord–Alta, Finnmark (Holmsen 1916)	Postglacial uplift has been estimated from levelling of shorelines in western Finnmark. The uplift shows negative anomalies diverging from the regional trend in the order of 5 m in the Øksfjord area. This effect was attributed to the gabbro massifs within the Seiland Igneous Province.	The interpretation is hampered by poor age control on the formation of the shorelines.	C7
3	Masi–Iešjav'ri area, Finnmark (Olesen 1988, Solli 1988, Muir Wood 1989a, Olesen et al. 1992a,b, Bungum and Lindholm 1997, Roberts et al. 1997, Olsen et al. 1999, Dehls et al. 2000a, Sletten 2000)	The NE–SW-trending postglacial Stuoragurra Fault (SF) extends for 80 km in the Masi–Iešjav'ri area in the Precambrian of Finnmarksvidda. The fault is manifested on the surface as a fault scarp up to 7 m high and is situated within the regional, Proterozoic, Mierujavri–Sværholt Fault Zone. The SF is a southeasterly dipping reverse fault. A c. 1m-thick zone containing several thinner (a few cm wide) zones of fault gouge represents the actual fault surface. The 21 January 1996 earthquake (M 4.0) in the Masi area was most likely located along the SF at a depth of c. 10 km.	The age of the SF is constrained in that it cross-cuts glaciofluvial deposits northeast of Iešjav'ri and an esker northeast of Masi. Thus, it formed after the deglaciation (c. 9,300 yr BP).	A1
4	Lyngen, Troms (Holmsen 1916)	Holocene uplift was assessed from levelling of shorelines in northern Troms. Negative uplift anomalies in the order of 5 m were ascribed to gabbro massifs within the Lyngen Ophiolite.	The interpretation is hampered by poor age control on the formation of the shorelines.	C7
5	Nordmannvik-dalen, Kåfjord, Troms (Tolgensbakk and Sollid 1988, Sollid and Tolgensbakk 1988, Blikra et al. 2006, Dehls et al. 2000a, Osmundsen et al. 2009)	NW–SE-trending postglacial faults in the Kåfjord area, North Troms. Normal faults dipping 30–50º to the northeast (Dehls et al. 2000a). The height and length of the main escarpment is approximately 1 m and 2 km, respectively.	The fault is subparallel to the Nordmannvikdalen valley. The slope of the terrain is 10–12º and the elevation difference between the fault scarp and valley bottom is 150–200 m. According to Varnes et al. (1989) gravity-induced sliding is less likely to occur when the elevation difference is less than 300 m. We therefore favour a tectonic origin for the fault.	A1
6	Balsfjord–Lyngen area, Troms (Blikra et al. 2006)	A distinct concentration of gravitational faults and slope failures may indicate a large-magnitude, prehistoric earthquake. Several hundred, large rock-slope failures and landslides were triggered during this event.	The slope failures in Troms county seem to be old (during and shortly after the last deglaciation), and are most likely related to the enhanced seismic activity shortly after the deglaciation.	B1, 2
7	Astafjorden–Grytøya area, Troms (Blikra et al. 2006)	A relatively high concentration of rock avalanches.	The number of rock avalanches is not as high as in the Balsfjord–Lyngen area farther north, but is much higher than in the Senja area where the topography is considerably rougher.	C1, 2
8	Kvasshaugen, between the valleys of Beiardalen and Gråtådalen, Nordland (Grønlie 1939, Muir Wood 1993)	NNE–SSW-trending clefts occur along an approximately 5 km-long NNE–SSW-trending zone. These clefts are up to 20 m wide and 10 m deep and the eastern sides are locally down-faulted.	The faults may be classified as sackung features (Varnes et al. 1989) due to gravity spreading of the 500 m-high ridge along Gråtåhaugen, Kvasshaugen and Monsfjellet (Olesen et al. 2004). The initiation of movements, however, may have been triggered by large earthquakes.	B1, 2
9	Ranafjord area, Nordland (Helzen 1834, Grønlie 1923, Muir Wood 1989b, Bakkelid 1990,2001, Olesen et al. 1994, 1995, Hicks et al. 2000b, Olsen 1998, 2000)	The Båsmoen fault consists of SSE-dipping (40–70º) fault segments within a 2 km-wide and 50 km-long zone. There is evidence for anomalous land uplift along the Båsmoen fault at the locations Utskarpen, Straumbotn and Båsmoen on the northern shore of Ranafjord and Hemnesberget. Numerous liquefaction structures have been observed in the area. The fault bears resemblance to the postglacial faults reported from the Lapland area of northern Fennoscandia.	No conclusive evidence has yet been found for postglacial movements along specific fault scarps. Trenching of the fault scarp indicates a 40 cm offset along the Båsmoen fault (Olsen 2000). A new seismic miniarray has registered numerous minor earthquakes in the outer Ranafjord area. They do not, however, seem to be attributed to the Båsmoen fault (Hicks et al. 2000b).	B1, 3

Loc.No.	Location and reference	Observation	Comment	Grade/ type
10	Nesna islands, (Handnesøya, Hugla and Tomma), Nesna, Nordland (Bakkelid 1990, 2001, Olesen et al. 1994, 1995, 2012b)	N–S-trending, steeply dipping fractures and faults occur on Handnesøya. There seems to be a vertical offset of the bedrock surface across these structures. The foliation of the mica schist is subparallel to the bedrock surface. Field inspection revealed that the features are probably of erosional origin since the scarps seem to have been sculptured by the moving inland ice. Monitoring of the seismicity in the area, however, has shown that more than 20 earthquakes occurred along one of these fault zones in 1998, indicating that the fault is active at depth (Hicks et al. 2000b). The observed uplift of acorn barnacle and bladder wrack marks at two different locations on the neighbouring island Hugla deviates 1–2 mm yr^{-1} from the regional trend. The benchmarks were established in 1894 and remeasured in 1990. GPS campaign measurements in 1999 and 2008 by the Norwegian Mapping Authority (Kartverket) indicate that the bench marks to the west of the earthquakes have moved c. 1 mm yr^{-1} to the NW and W relative to the stations to the east of the earthquake swarms (Olesen et al. 2012b). Dehls et al. (2002) reported irregular subsidence in the order of 0.5 mm yr^{-1} from InSAR permanent scatterer data during the period 1992–2000.	The ice has most likely plucked blocks from the bedrock along steeply dipping N–S-trending fractures (Olesen et al. 2004). The observed uplift at the two locations on Hugla is 0.0 and 0.5 mm yr^{-1}. The benchmark with the zero uplift may, however, have been moved from its original position (S. Bakkelid, pers. comm. 2000). Fault plane solutions from some of the frequent earthquakes in the area reveal extensional faulting consistent with the observed subsidence (Hicks et al. 2000b). The consistent pattern of present-day extension and subsidence from four different methods is a strong indication of active deformation in the outer Ranafjorden area, although no direct active fault scarp has been detected at the surface.	A1
11	Klubbsteinen, Nord-Flatanger, Nord-Trøndelag (Olsen and Sveian, 1994, Olsen 1998)	A thick deposit of fine to medium sand with clast-supported conglo-meratic character is recorded in the c. 4 m-high sections of a sand pit near the intersection of two old fault/fissure lineaments. The sand is truncated on the top and overlain by a subhorizontal, bedded, gravelly sand of c. 1.0 m thickness. The two sand units comprise the material of a strand terrace which corresponds to the Tapes maximum sea level.	The observed clast-supported conglomeratic character of the sand resembles the compositions of similar sands recorded at c. 10 other sites in Mid and North Norway. Earthquakes seem to be a likely cause of the production of these characteristics, and are, in fact, in some cases, e.g., at Klubbsteinen (named Sitter in Olsen 1998), the only reasonable triggering mechanism for this phenomenon. The conglomeratic feature has clearly been developed quite suddenly, some time after the original subhorizontal and alternating layering of fine and medium sand. The age of these earthquake-related features must be older than the regression from the maximum Tapes sea level, but younger than the culmination of the Tapes transgression, i.e., c. 7000–7500 ^{14}C yr BP (Sveian and Olsen 1984).	B1
12	Tjellefonna, Langfjorden, Nesset, Møre & Romsdal (Morsing 1757, Bungum and Lindholm 2007, Redfield and Osmundsen 2009)	Morsing (1757) concluded that the Tjellefonna and Silset rock avalanches on 23 February 1756 in Møre & Romsdal county were caused by an earthquake (Bungum and Lindholm 2007, Redfield and Osmundsen 2009). The Tjellefonna rock avalanche created a 40 m-high tsunami that caused catastrophic damage to the settlements along Langfjorden and Eresfjorden and killed a total of 32 people. Redfield and Osmundsen (2009) suggest that an earthquake on a nearby fault caused the already weakened Tjelle hillside to fail.	It is possible that the Tjellefonna rock slide was triggered by an earthquake since the historical records suggest a contemporaneous rockslide at Silset located 15 km farther north. Morsing (1757) described a long rumbling noise just before the Tjellefonna rock avalanche plunged into the Langfjorden. He interpreted this in terms of an earthquake.	C1, 2
13	Berill, Rauma, Møre og Romsdal (Anda et al. 2002)	The NNE–SSW-trending Berill Fault is 2.5 km long and has an offset of 2–4 m. The reverse fault dips to the west. It truncates well-defined colluvial fans and was formed after the Younger Dryas period. The fault is little modified by avalanche processes, suggesting that it originated during the second half of the Holocene. There are fault scarps up to 15 km in length but dating of these sections is lacking.	The fault represents a reactivation of an older fault zone (cohesive cataclasite) and it occurs in a zone with a large number of rock-slope failures. This reverse fault appears to be part of a sackung structure (Savage and Varnes 1987). The open clefts with upward-facing scarps along the mountain ridge in the hanging-wall block of the reverse fault are typical features of gravity-induced sackung structures. The low offset/length ratio (1/500) of the fault also points to a nontectonic origin. A nearby large-magnitude earthquake may have triggered the collapse structure.	B2
14	Oterøya–Øtrefjellet, Haram, Møre og Romsdal (Braathen et al. 2004, Blikra et al. 2006)	Concentrations of rock-slope failures and collapsed bedrock. This includes a 2 km-long, N–S-oriented mountain ridge on Øtrefjellet. It is situated 100–300 m above the surrounding terrain and is heavily fractured. Crushed or collapsed bedrock occurs locally within a 500 m-wide zone. The slopes of the ridge are too gentle to cause any gravity sliding (Braathen et al. 2004, Blikra et al. 2006). An earthquake could have provided the necessary energy for triggering the failure.	An alternative mechanism is that of processes related to permafrost conditions during colder phases after the Weichselian maximum. The ridge is situated 200–300 m below the distinct 'weathering zone' of the region, thought to represent the ice limit during the Weichselian maximum (Anda et al. 2000).	B1, 2

Loc.No.	Location and reference	Observation	Comment	Grade/ type
15	Ørsta–Vanylven, Møre og Romsdal (Bøe et al. 2004, Blikra et al. 2006)	Several large rock-slope failures and regional slide events in several fjords indicate that earthquakes may have played a role as a triggering mechanism.	There are indications of three regional slide events, one shortly after the deglaciation, one at 8000 and one at 2000 calendar years BP. The 8000 BP event is interpreted to be related to the tsunami generated by the Storegga slide.	C3
16	The Norwegian coast between Stadt and Vesterålen (Holtedahl 1984, 1998, Møller 1985, Sjöberg 1988)	Several authors have observed that sea-formed caves are located at a higher altitude than the postglacial marine shore level along the coast between Stadt and Vesterålen. They concluded that a long-term neotectonic uplift continued through the late Quaternary period. The sills of several caves are situated c. 30 m above the upper marine shoreline and the height of the cave opening varies between 30 and 50 m.	Olsen and Grøsfjeld (1999) have reported uplifted (40–90 m) Middle and Upper Weichselian marine sediments at several locations in Norway. These positions indicate a frequently fluctuating ice sheet during the interval 18–50 ka BP. Repeated rapid ice retreat following heavy ice loading could, to some extent, explain the location of sea-formed caves above the postglacial marine limit. It is, however, an open question if this model would imply sufficiently long time periods for the caves to have been formed by sea erosion. Deeply weathered fracture zones could have facilitated the formation of such caves.	C3
17	Aurland–Flåm, Sogn & Fjordane (Blikra et al. 2006)	A series of rock-slope failures, including an up to 4 km-long normal fault of probable gravitational origin. Dating of cores from the fjord suggests that large-scale rock-slope failures occurred shortly after the deglaciation.	It is still uncertain whether the normal fault is simply a gravitational feature, or if it may be linked to a tectonic structure at depth.	B1, 2
18	Geitura, Ulvik, Hordaland (Simonsen 1963, Blikra et al. 2006)	A large rock-avalanche on a fairly gentle slope. An earthquake has most likely triggered the avalanche.	This observation indicates the occurrence of at least one large-magnitude earthquake in the inner Hardanger area. However, postglacial fault scarps have not been found. Other large rock-avalanches have also been identified in the area (Blikra et al. 2006). Neotectonic activity in the Hardangerfjord area is supported by recent work on shoreline displacement by Helle et al. (2000).	B1
19	Finse–Geilo area, Hordaland–Buskerud (Anundsen et al. submitted)	Anomalous uplift from repeated levelling.	A careful analysis of the levelling methods is pending.	C8
20	Fjøsanger, Hordaland (Mangerud et al. 1981)	A considerable, long-term, neotectonic uplift (10–40 m) of the Bergen area during the last 125,000 years is based on investigations of marine sediments from the Eemian interglacial.	The ice melted more rapidly at the end of the Saalian than at the end of the Weichselian (Ehlers 1990). This difference in the rate of deglaciation may explain the occurrence of marine Eemian sediments at elevated positions on Fjøsanger.	C
21	Yrkje area (Anundsen and Fjeldskaar 1983, Anundsen 1985)	A 7–10 m offset (since 10,400 BP) of the Younger Dryas transgression level across a NE–SW-trending fault. The observation is based on a study of the marine isolation of six basins and one of these basins shows anomalous uplift.	The offset is only observed at one location. No further work has been undertaken to study other lake basins in this area. From a limited set of dates and cores, the fault explanation for the apparent variation in isolation levels is not unique (Muir Wood 1993).	C7
22	Bø, Karmøy, Rogaland (Sejrup 1987)	The location of Eemian sediments at an estimated altitude of 15–45 m above the Eemian sea level is applied to deduce a long-term uplift of the Karmøy area during the last 125,000 years.	See comments on location no. 42 at Fjøsanger.	C
23	Egersund–Flekkefjord area, Rogaland (Anundsen 1989, Anundsen et al. submitted)	A subsidence of 2–2.5 mm yr^{-1} has been recorded in the Egersund Anorthosite–Gabbro Province. The zone of maximum subsidence coincides with a maximum gravity anomaly in the area.	A follow-up geodetic study is needed to carry out a better evaluation of this claim.	C8
24	Haukeligrend, Vinje, Telemark (Anundsen et al. submitted)	Anomalous subsidence from repeated levelling.	A careful analysis of the levelling methods is pending.	C8
25	Storegga (Evans et al. 1996, Bryn et al. 1998)	Postglacial N–S-trending faults and a graben, up to 150 m wide, reaching the sea bed or coming to within a few metres of it. Throws of up to 4 m have been recorded. The length of the composite structure is more than 5 km.	There are no regional deep-seated faults below the fault scarps. There is an abundance of pockmarks in the area. The faulting may be related to gas escape, as suggested by Fulop (1998). The faults bear a resemblance to the structures observed by Hovland (1983) at the western margin of the Norwegian Channel, which have also been attributed to gas leakage. The features have possibly been triggered by earthquakes.	C, D
26	Faeroe–Shetland Escarpment, (Seabed Project, NORSAR 1999)	A 25–30 m-high offset in the Storegga slide deposits above the Faeroe–Shetland Escarpment.	The fault cannot be observed on any of the adjacent lines and must consequently be shorter than 20 km. The height/length ratio of the fault is 0.0012–0.0015, which is greater than for most other postglacial faults in Fennoscandia.	C, D

Loc.No.	Location and reference	Observation	Comment	Grade/ type
27	Troll–Fram area, North Sea (Riis 1998, Olesen et al. 1999)	WNW–ESE lineaments in the Early Pleistocene, interpreted in 3D surveys, coincide with Mesozoic faults reactivated in the Tertiary.	The lineaments are close to, or below, the seismic level of resolution. There is abundant faulting in the pre-Miocene rocks, related to escape of fluids and/ or gas.	C
28	Western slope of the Norwegian Trench, south of Kvitebjørn, offshore Øygarden (Hovland 1983)	A N–S-trending, normal fault-zone with 1–2 m offset. The fault-zone has a length of minimum 2 km and consists of 2–4 parallel faults commonly forming a subsided internal zone. The eastern zone is generally down-faulted. The faults were detected with a deep-towed boomer during the Statpipe route survey in 1981. The fault cuts soft, silty, cohesive clay.	The fault occurs in an area with abundant pockmarks and Hovland (1983) has suggested a genetic link between the two phenomena. A multibeam echo-sounding survey carried out by the Norwegian Mapping Authority in 1999 supports the conclusion reached by Hovland (1983). Release of gas does not, however, explain the 1–2 m offset of the sea floor. A tectonic cause cannot therefore be ruled out.	C1



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