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).

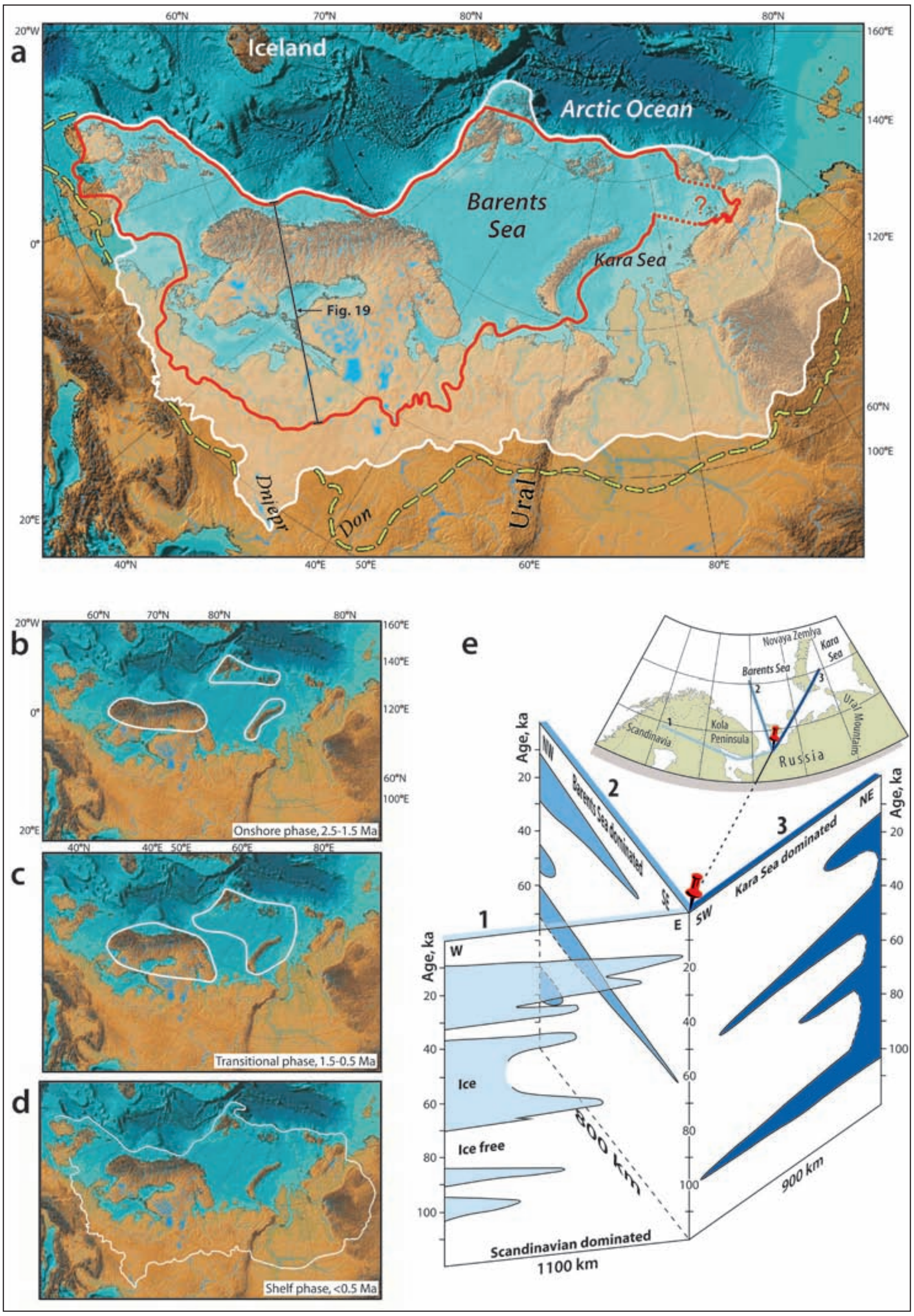


*Figure 1. Glaciations/glacial extension in northern Europe. (a) LGM (*red line*) and previous major glaciations (*white line*: Saalian and Elsterian glaciations;* yellow stippled line*: Quaternary maximum glacier extension), with coalescence between the Scandinavian–Fennoscandian, Barents Sea–Kara Sea 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.

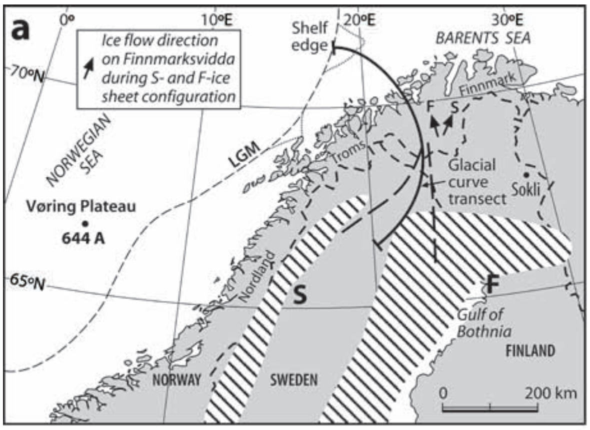
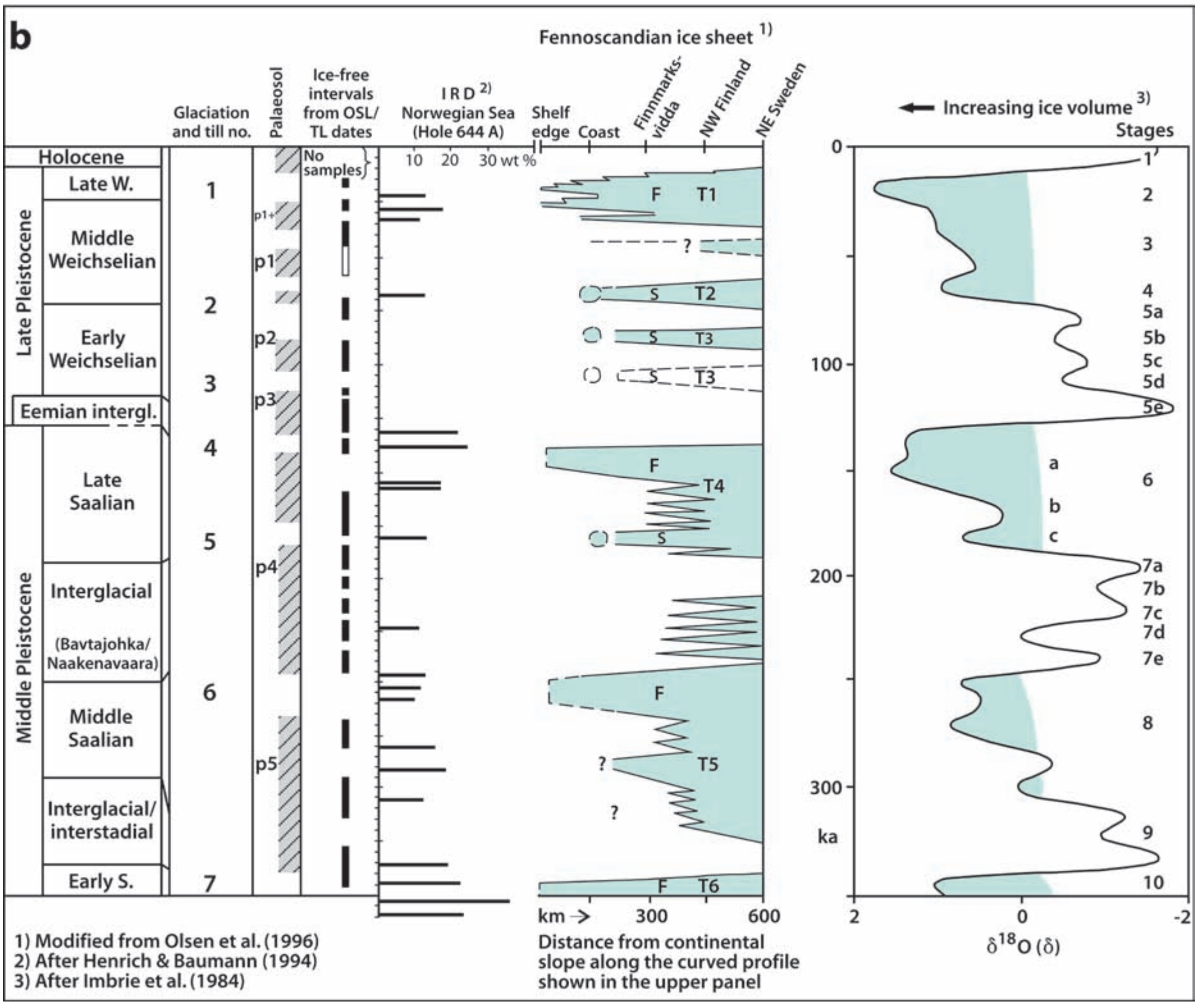


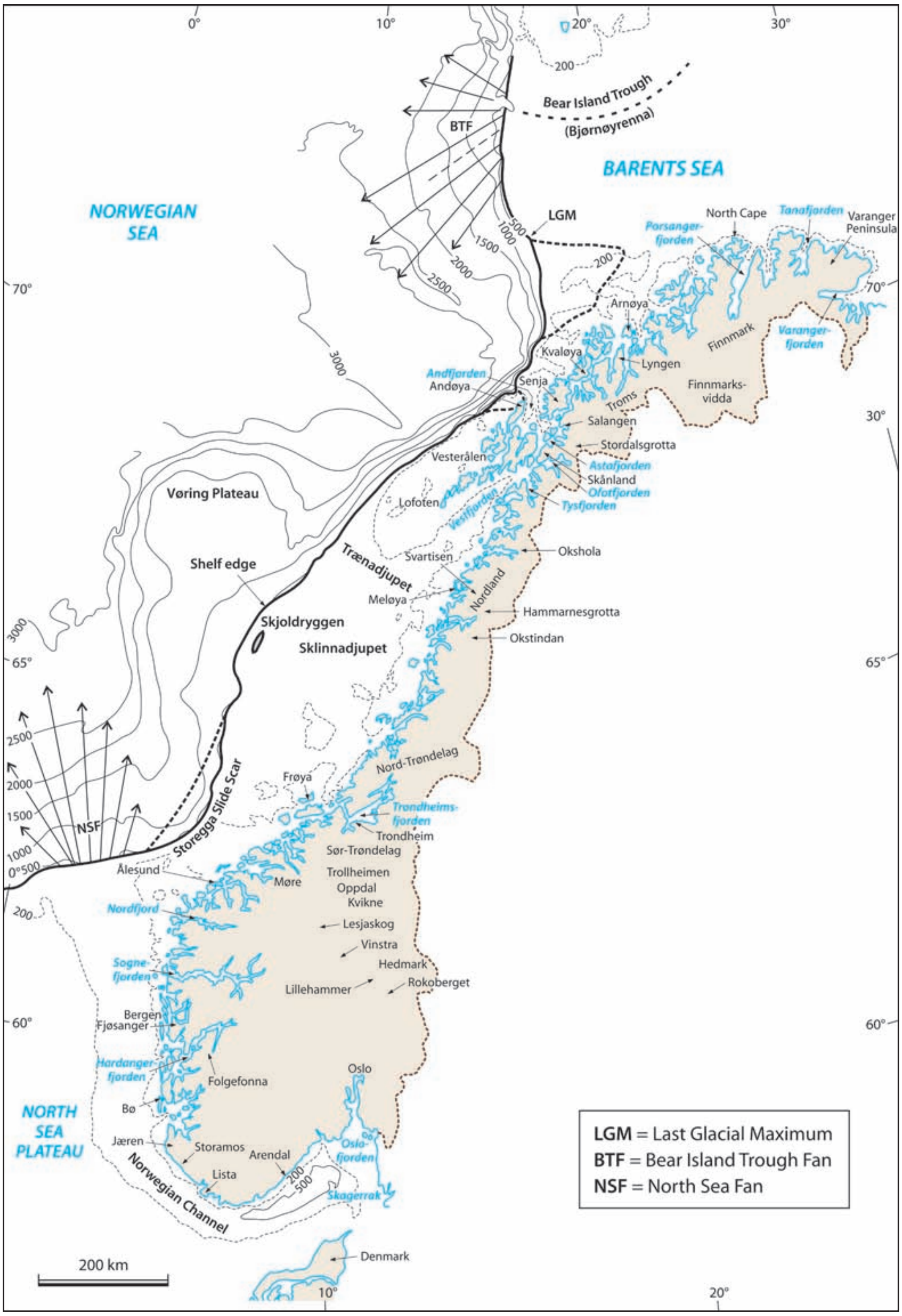
Figure 2. (a) Upper panel–*Northern and Central Fennoscandia with indication of positions for the ice dome -/ ice shed areas (shaded zones) during a Fennoscandian ice sheet configuration (F), with a maximum style glacier extension; and a Scandinavian ice sheet configuration (S), with a smaller ice extension (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,

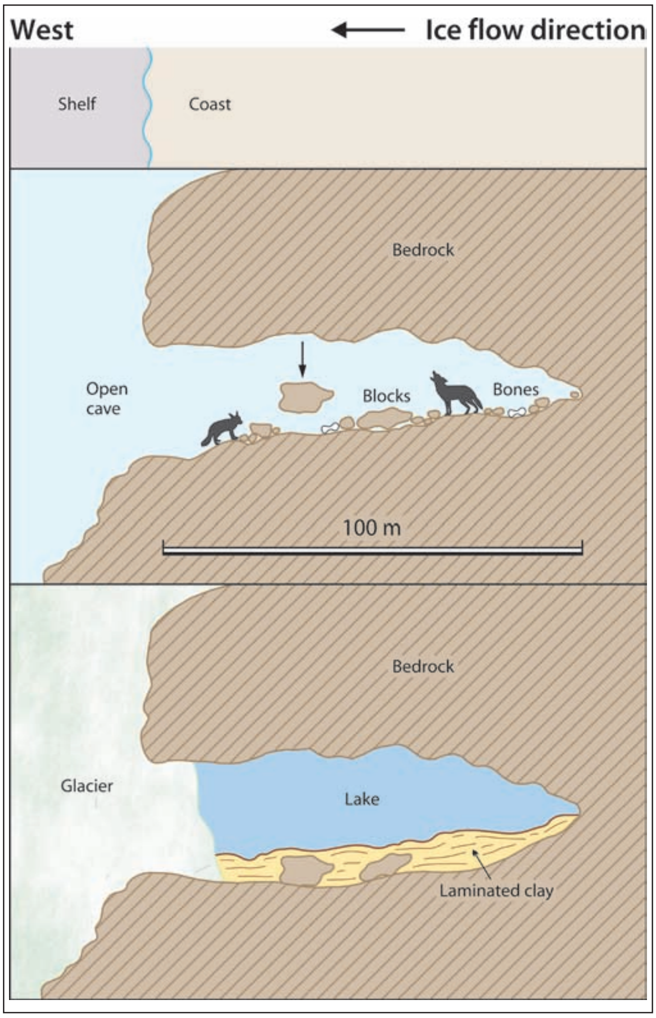


Figure 4 Conceptual cross sections of uplifted, marine caves with sediments in western Norway. The upper and lower sketches represent ice-free and ice-dammed conditions, respectively. Modified from Mangerud et al. (2003).

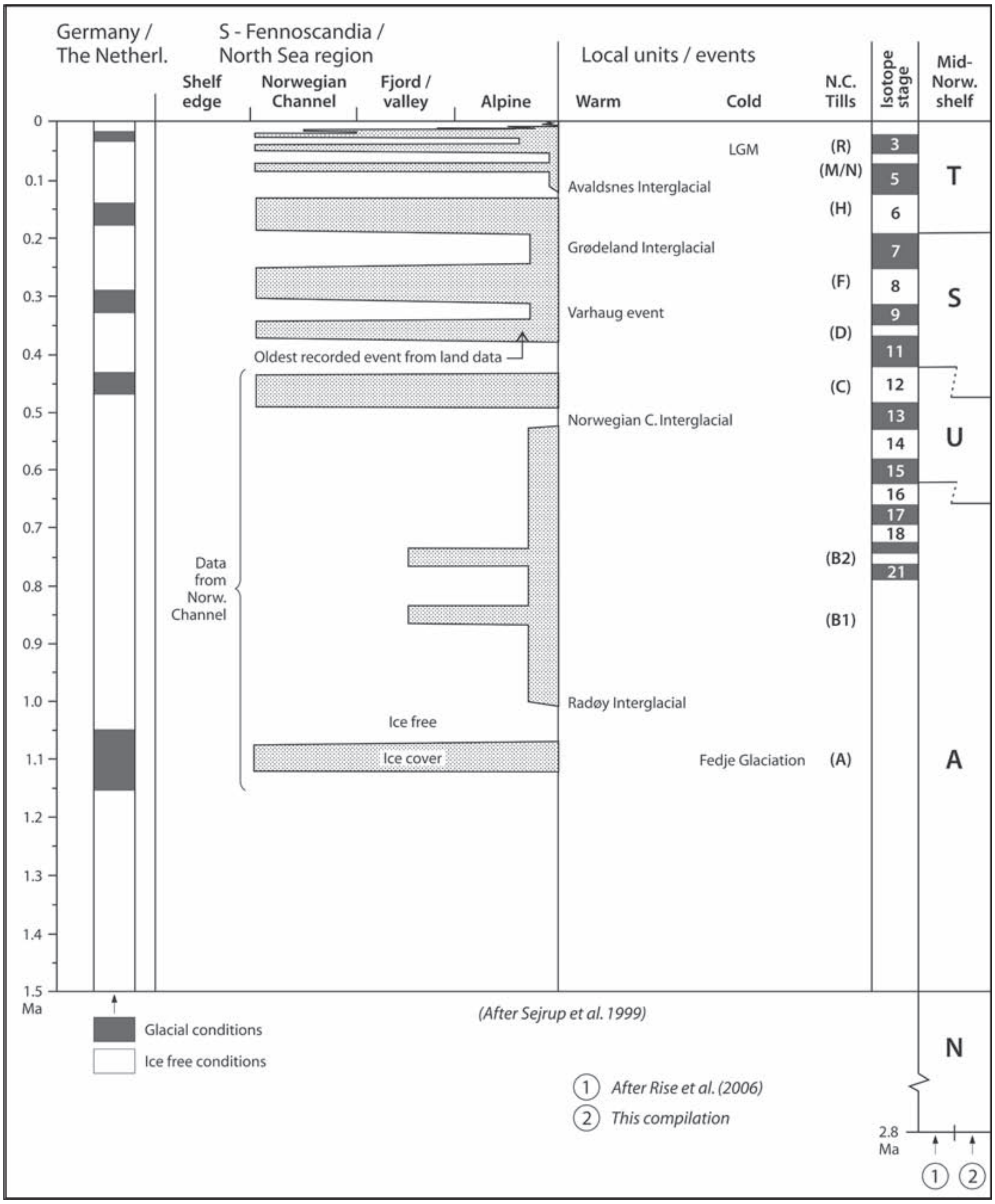


Figure 5. Glaciation curve for the Fennoscandian ice sheet, representing the southern part, including the North Sea region, during the last 1.1 million years. The stratigraphical positions of the till units A, B1, B2, C, D, F, H, M/N? (uncertain stratigraphical position) and R from the Norwegian Channel (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

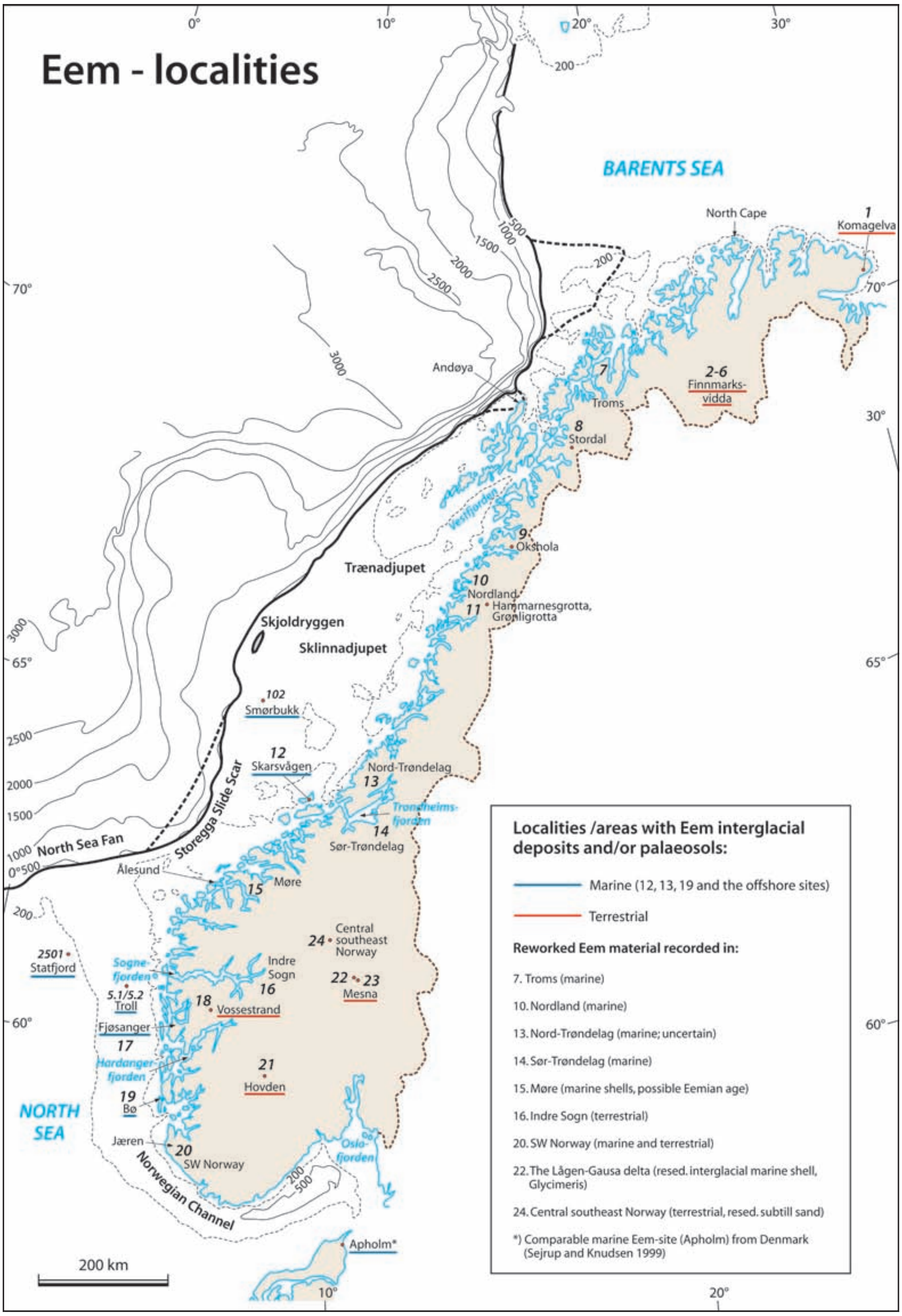


Figure 6. All reported Eemian sites on land in Norway and selected Eemian offshore sites. Onshore sites: 1–6 – Olsen et al. (1996a), Olsen (1998); 7 – Vorren et al. (1981); 8–9 – Lauritzen (1995); 10 – Olsen et al. (2001a), Olsen (2002); 11 – Linge et al. (2001); 12 – Aarseth (1990, 1995); 13–14 – Olsen 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

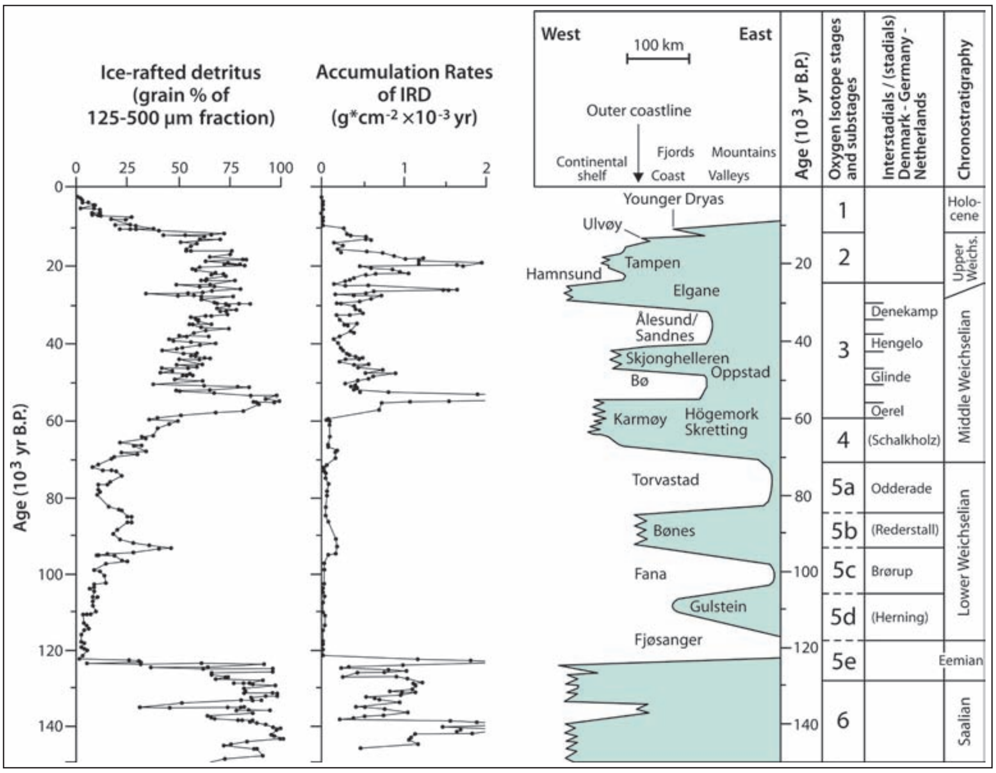


Figure 7. Comparison between IRD data from the Vøring Plateau west of the Norwegian coast and a glacial curve for southwest Scandinavia, west of the watershed and ice-divide zone. The time scale is in calendar years. Modified after Mangerud (2004) and Baumann et al. (1995).

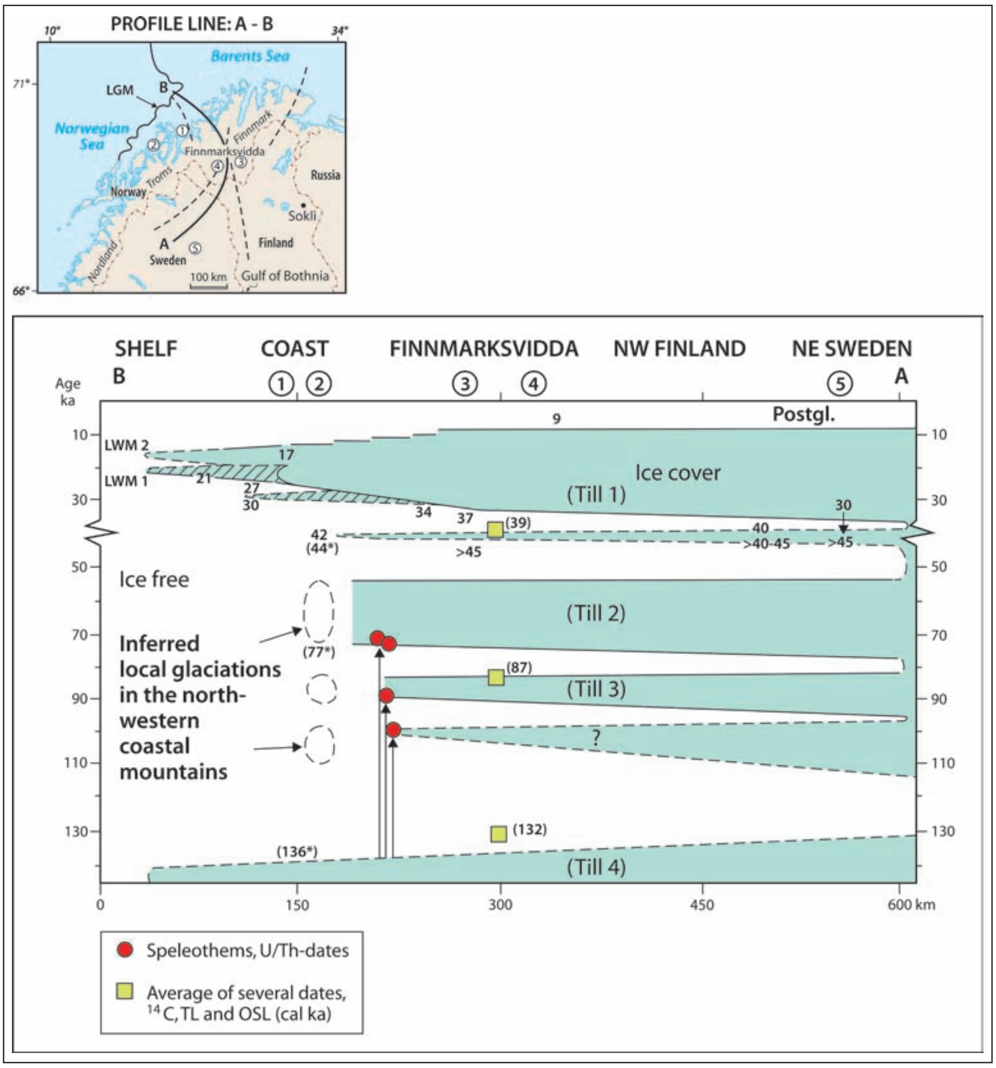


Figure 8. Glacial curve for the last 145,000 yr in northern Fennoscandia. Upper panel: *Location of profile line A–B and the stratigraphical site areas 1–5. The Last Glacial/Late Weichselian Maximum (LGM/LWM) limit at the shelf edge is indicated in the NW.* Lower panel: *Glacial curve for the last 145,000 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

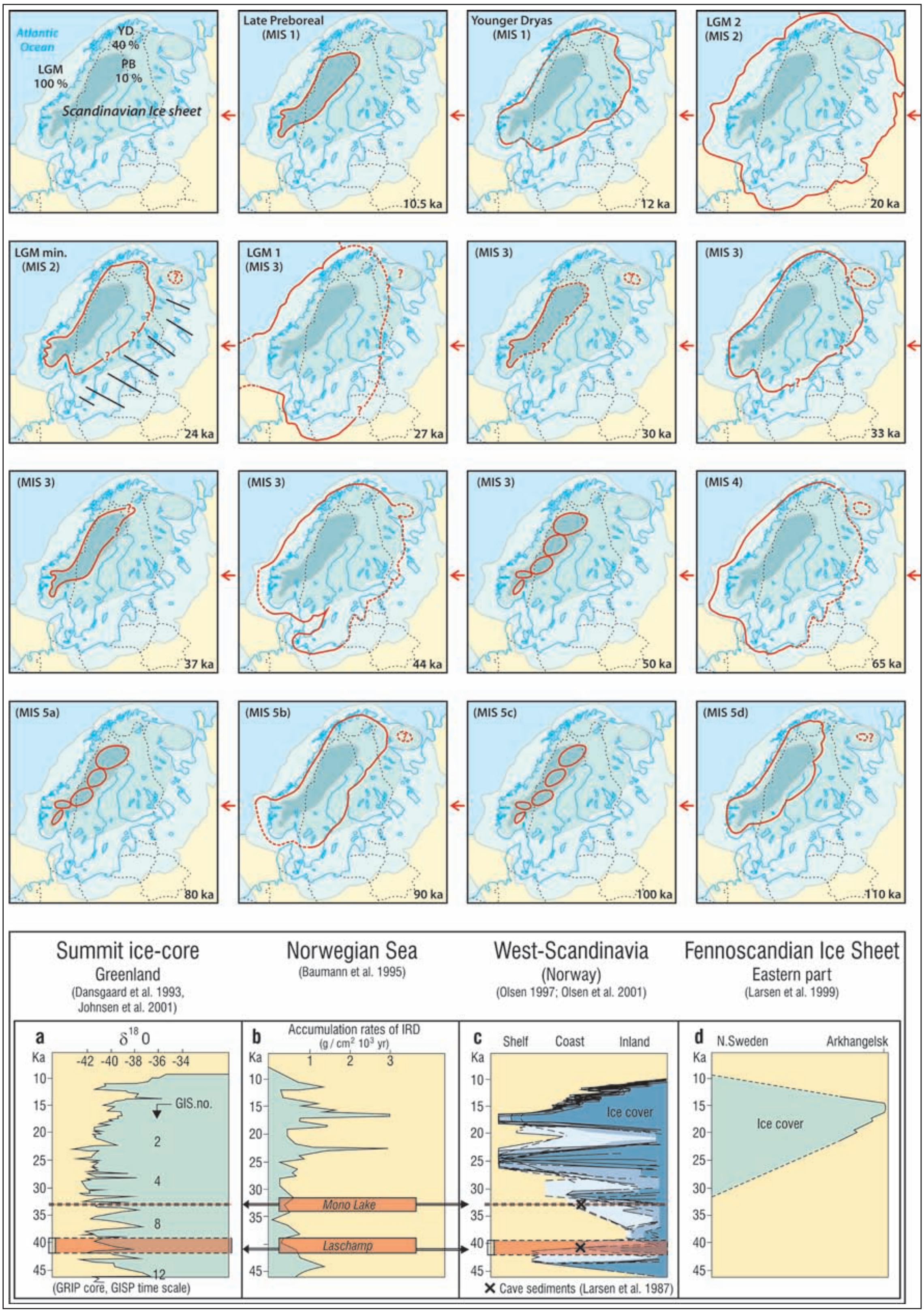


Figure 9. (a) Upper panel – Extension of the Scandinavian ice sheet during different stages of the Weichselian glaciation. The 16 maps are based on data from Lundqvist (1992), Larsen et al. (1999), Lunkka et al. (2001), Olsen et al. (2001a, b, c), Mangerud (2004) and Olsen (2006). The LGM ice extension 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 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).

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).

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

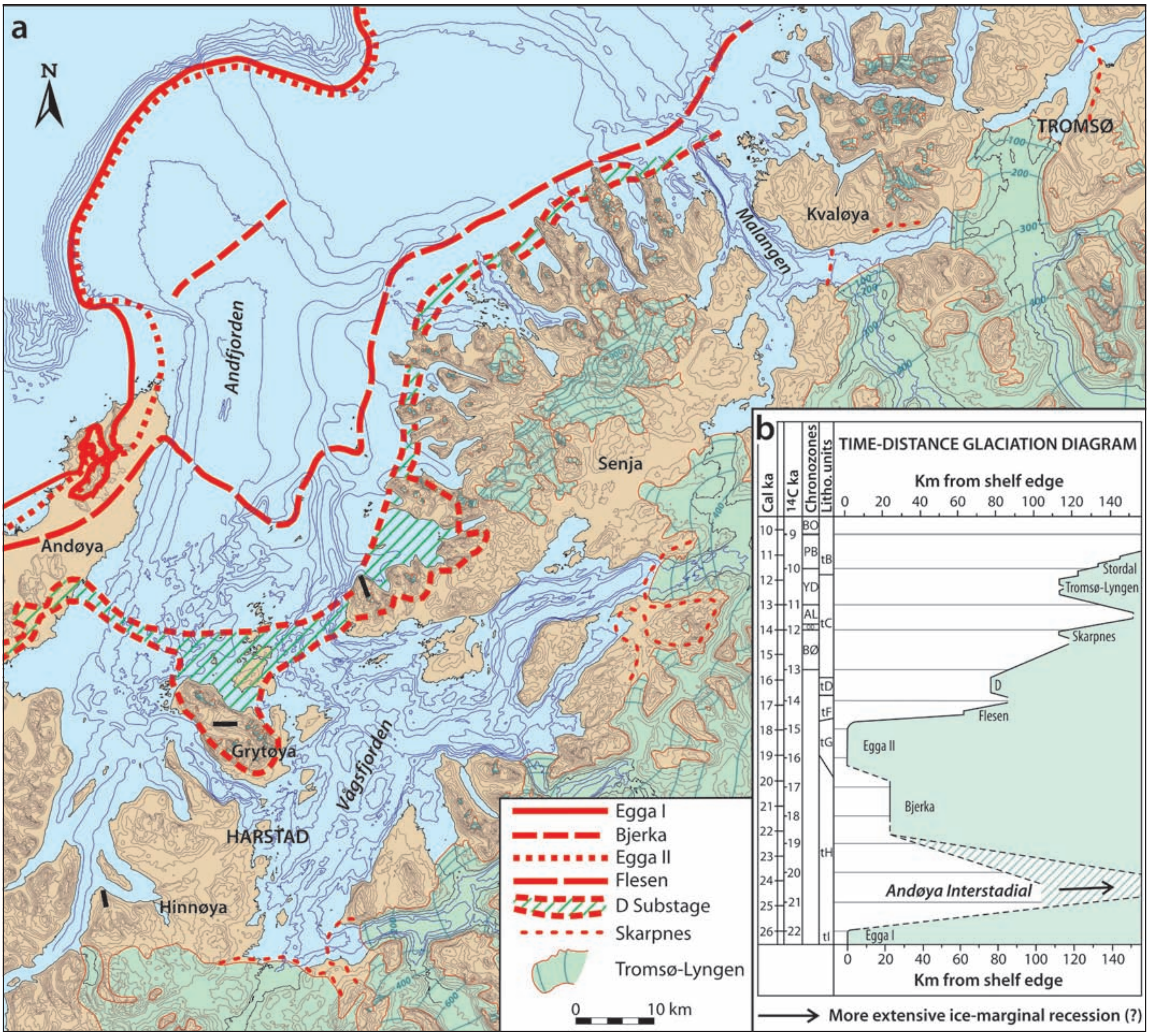


Figure 11. Late Weichselian ice margins (a) and glacial (time–distance) curve (b) showing the ice-front variations near Andøya island, northern Norway. Modified from Vorren and Plassen (2002). The ice margins indicated at Hinnøya in the southwest are after Sveian (2004) and Bergstrøm et al. (2005). Lateral 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,

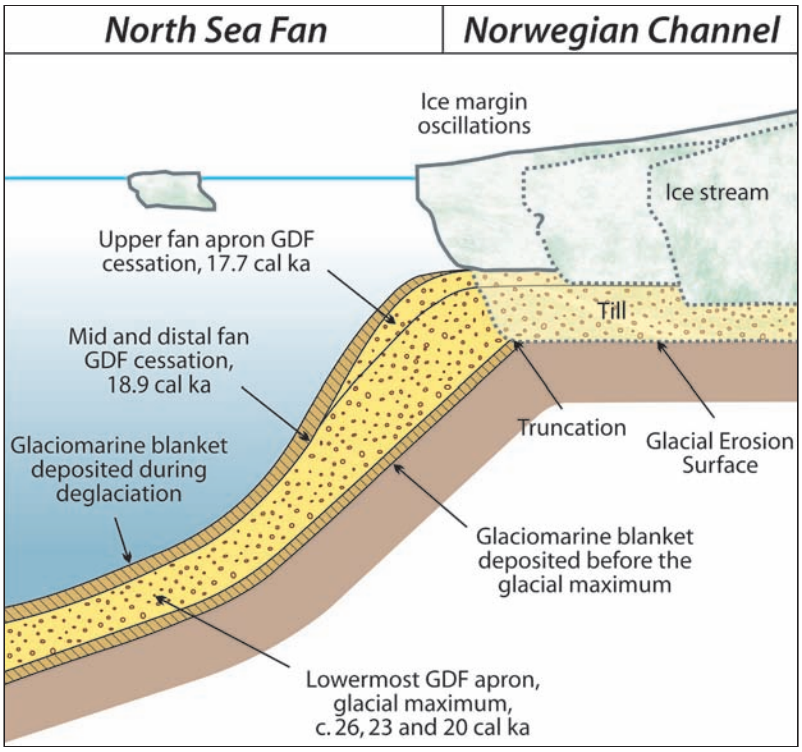


Figure 12. Schematic diagram showing the development of the Norwegian Channel trough-mouth fan during the Late Weichselian glacial maximum. GDF=glacial debris flow. After Mangerud (2004), slightly modified from King et al. (1998), with age assignments for the Last Glacial Maximum from Nygård et al. (2007).

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 resedimented 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

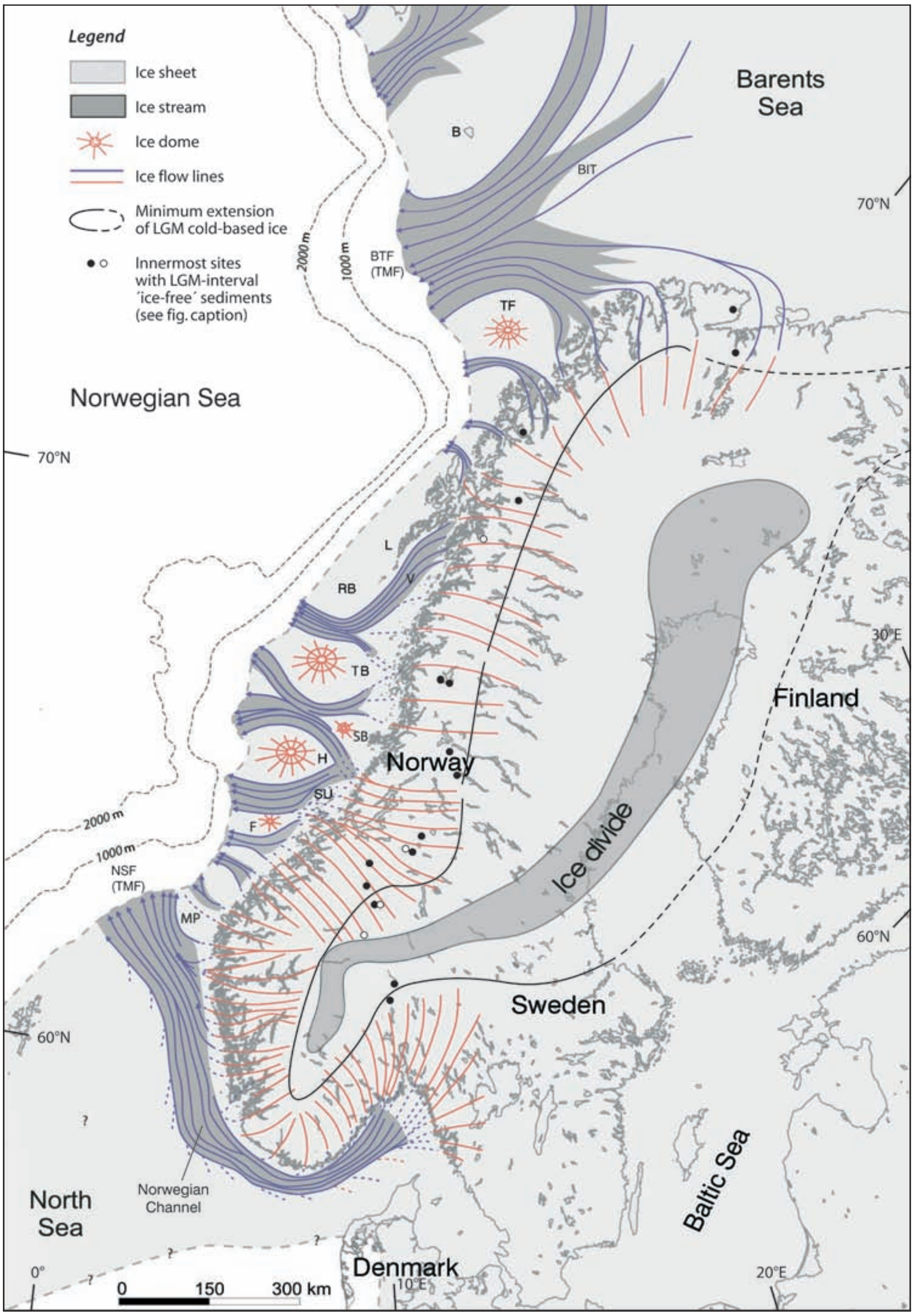


Figure 13. Interpreted ice-flow model during Late Weichselian glacial maximum, from Ottesen et al. (2005a). Minimum areas (discontinuous/stippled line) with ice frozen to the ground (after Kleman et al. 1997) and innermost locations of sites with 'ice-free' sediments from ice marginal retreat during 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.

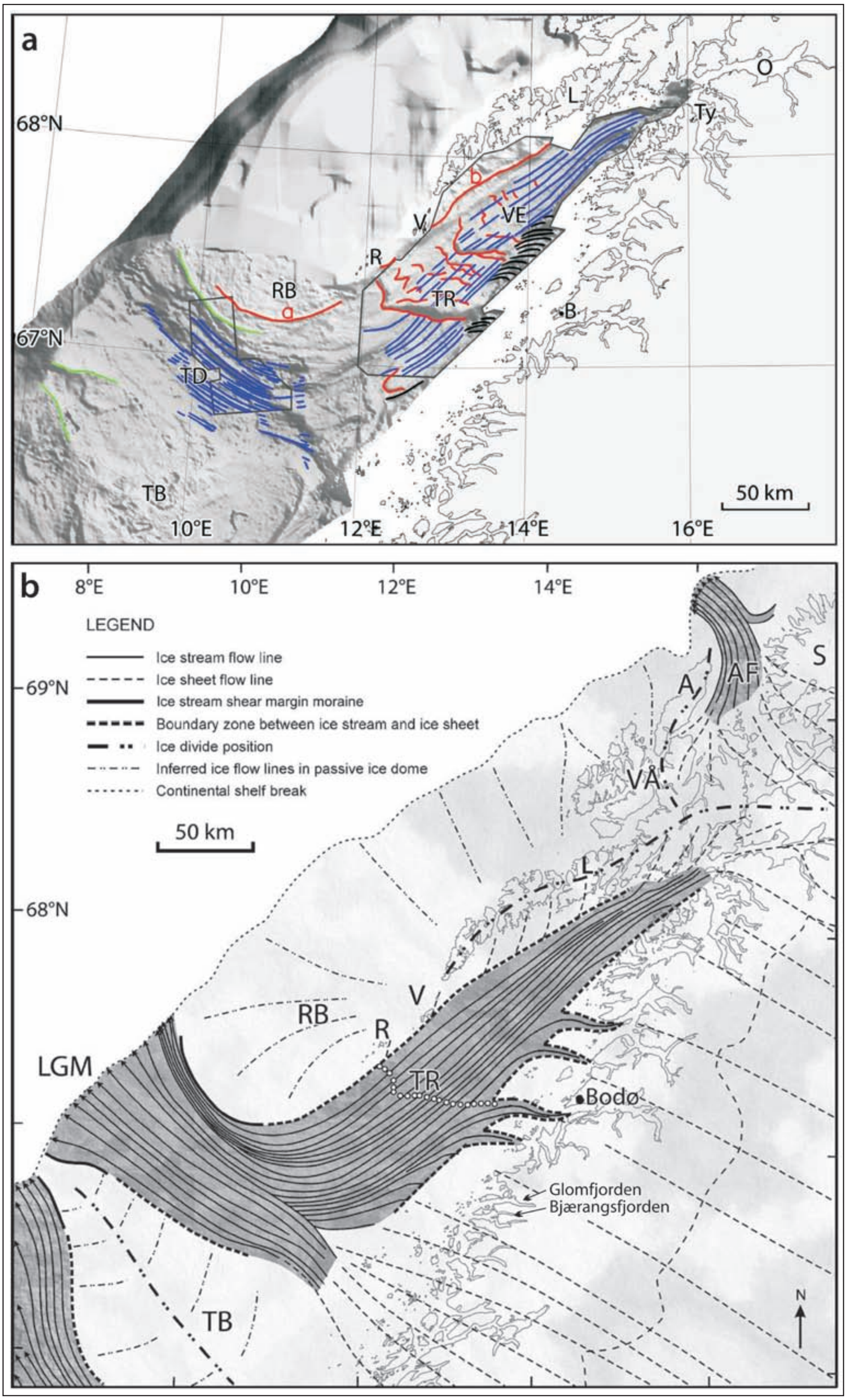


Figure 14. (a) Sea-floor morphology of the Vestfjorden/Trænadjupet with megascale glacial lineation (MSGL) indicating fast ice flow, and (b) Ice flow model of the Lofoten/Vesterålen area and adjacent shelf. Modified from Ottesen et al. (2005b). Legend of letters: O=Ofotfjorden, Ty=Tysfjorden, L=Lofoten, 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/

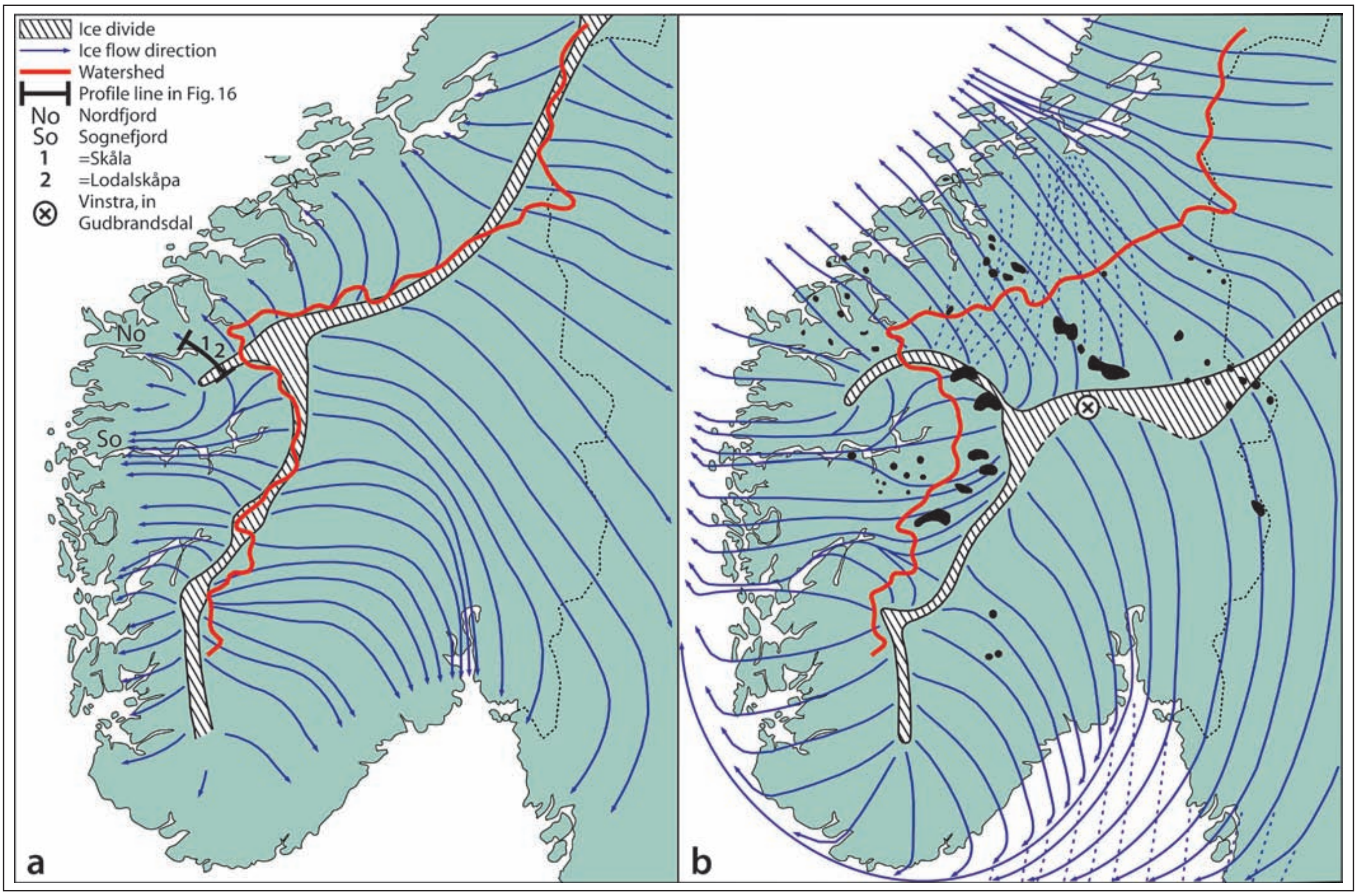


Figure 15. The main flow lines and divides of the ice sheet in southern Norway during small and moderate ice extension (a, left) and the LGM (b, right) are indicated. Modified from Vorren (1979). Blockfields (black spots) are also indicated (map b). Modified from Thoresen (1990). The present authors 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.

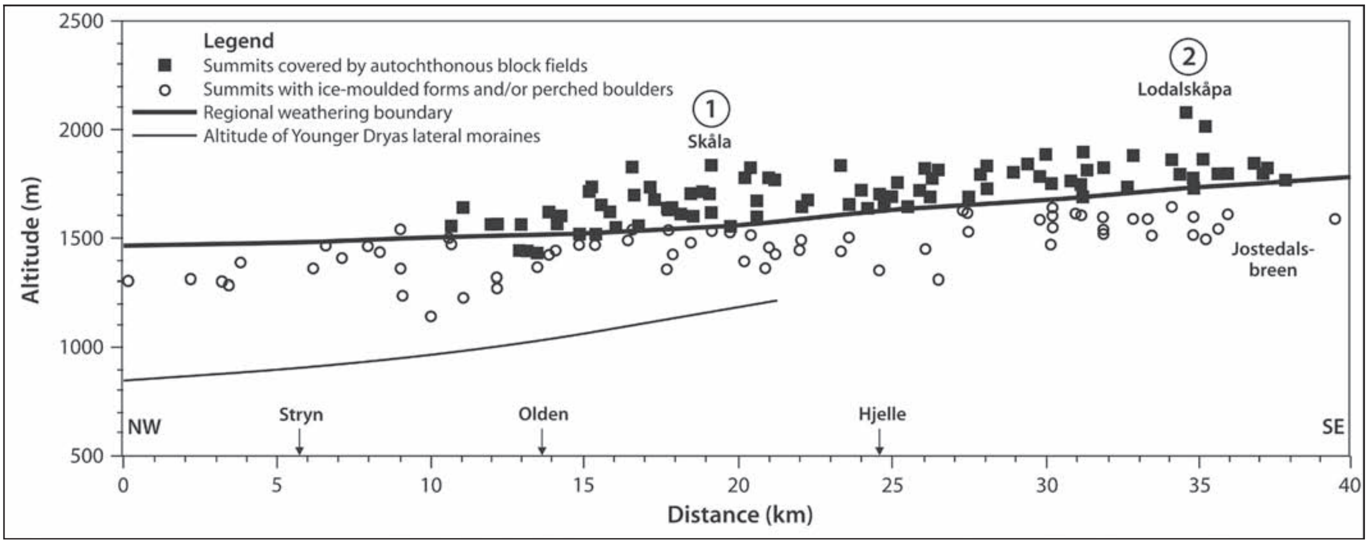


Figure 16. The distribution of summits, with and without block fields are plotted in a NW–SE cross section across inner Nordfjord (see Figure 15). The altitude of the Younger Dryas moraines is also shown. Modified from Mangerud (2004), updated from Brook et al. (1996). The lower boundary of the blockfields 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

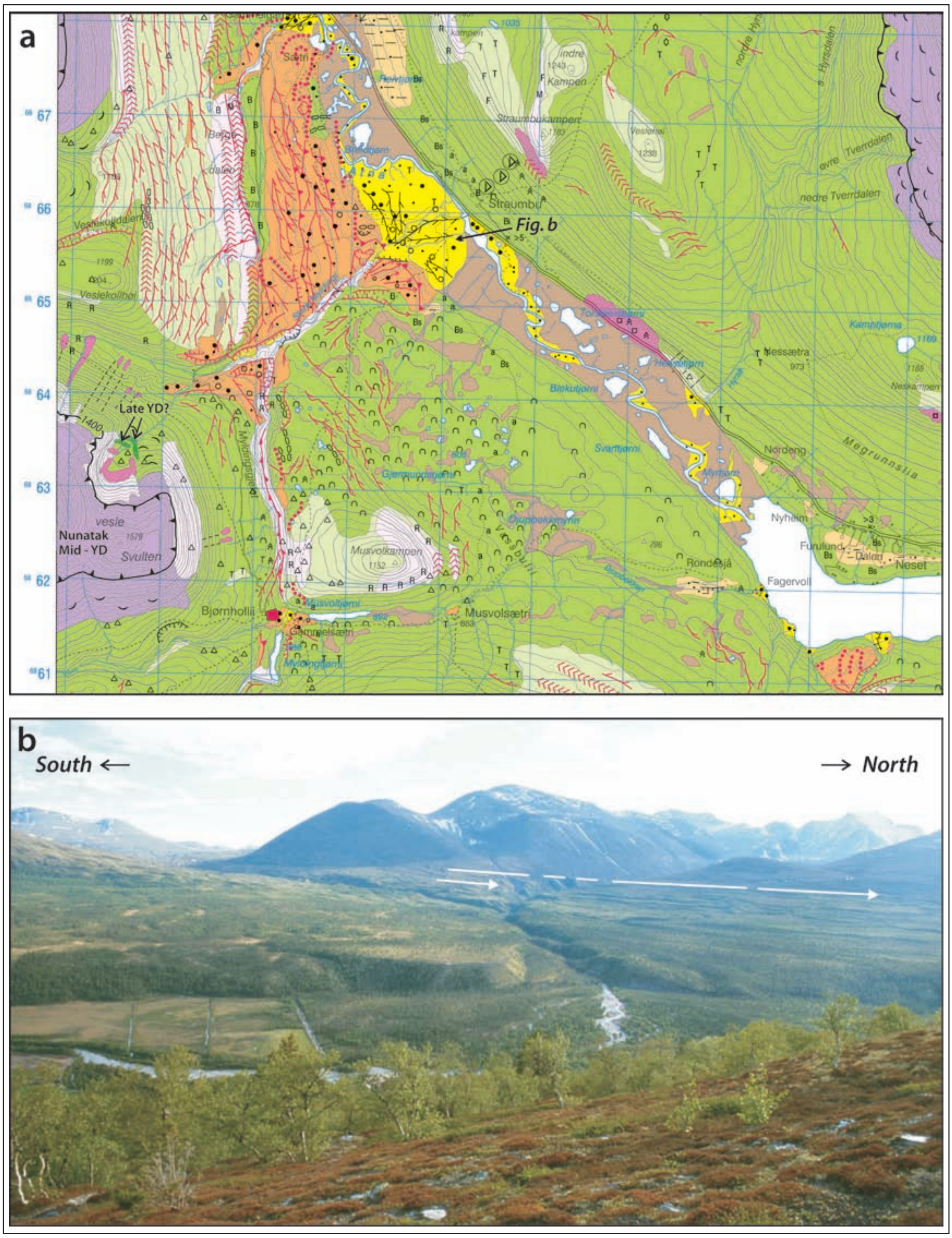


Figure 17. (a) Part of Quaternary map Atnsjøen 1818 IV (1:50,000). The view direction for (b) is marked. Size of each UTM coordinate net square is 1km x 1km. After Follestad (2006a). (b) Lateral meltwater channels from inland southeast Norway. Arrows indicate drainage direction towards the north. Photograph: 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

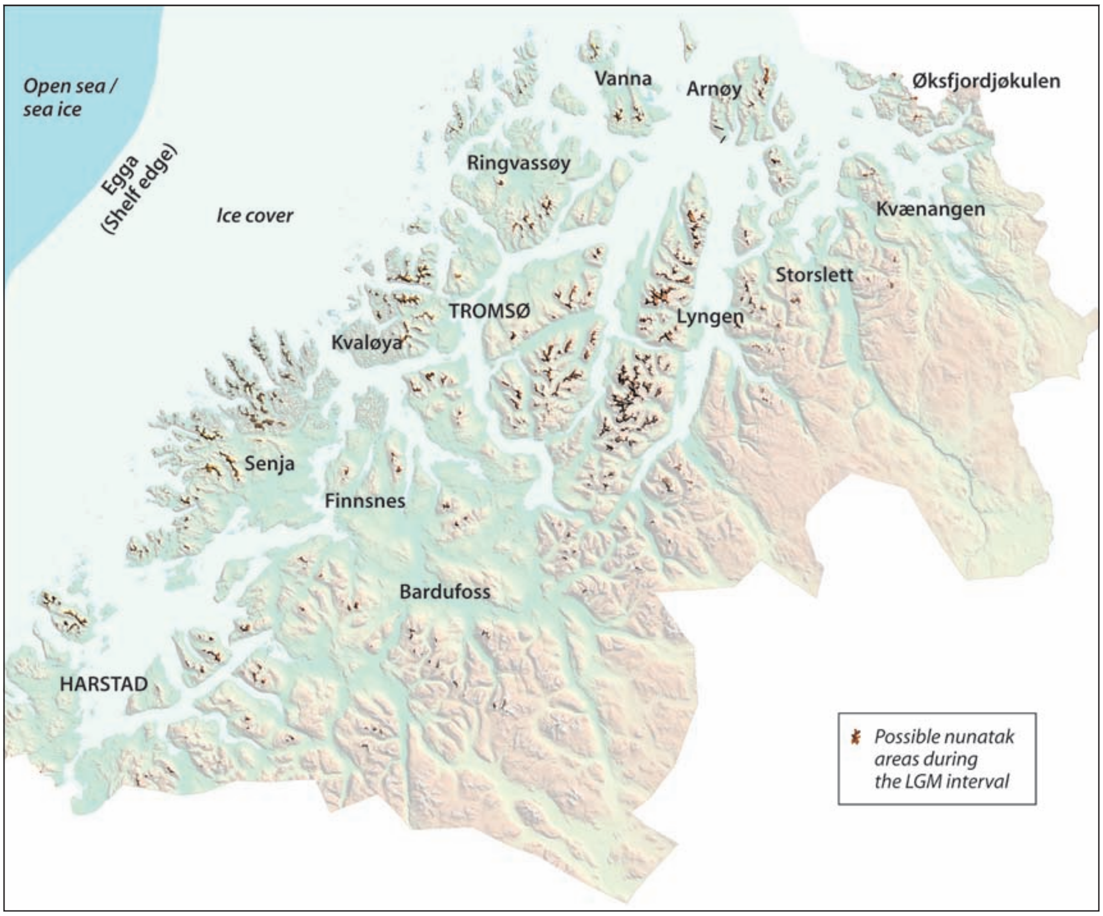


Figure 18. Location of high peaks (brownish shading), e.g., >800 m a.s.l. at Senja, which may have occurred as nunataks and penetrated the LGM ice surface along the coast, similar as in several other high mountain peak areas (>1000–1500 m a.s.l.) in the fjord and inland areas of Troms. Modified from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

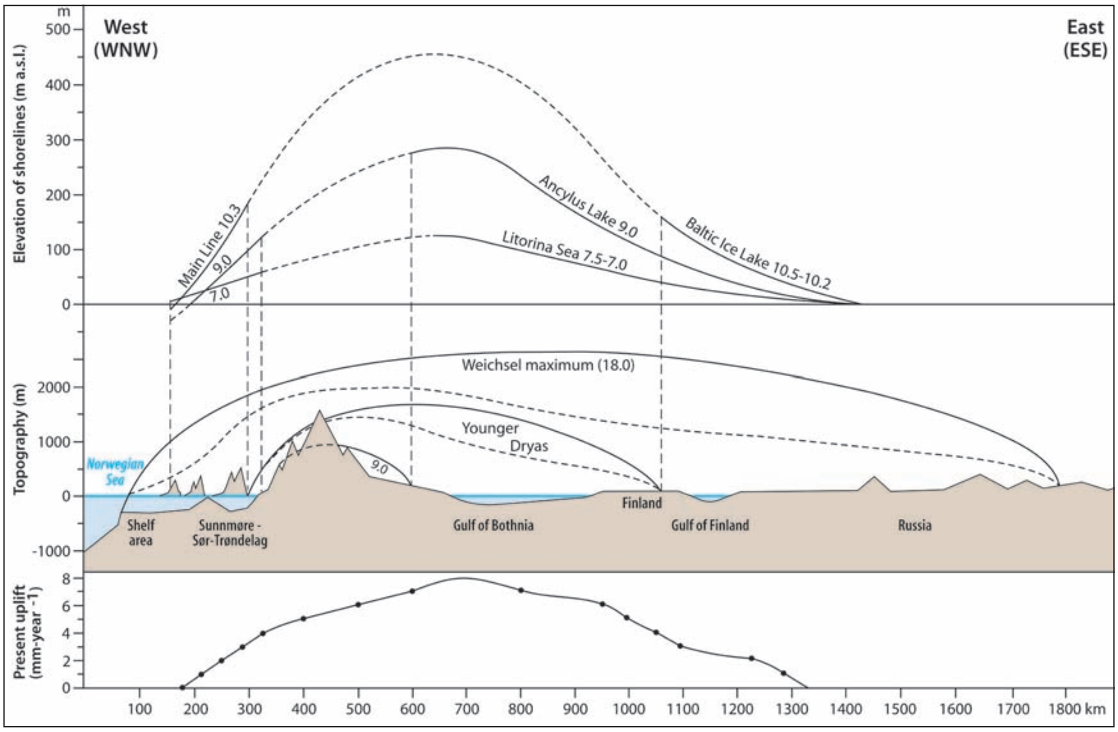


Figure 19. Profiles across Central Fennoscandia, generalised and modified from various sources, including Svendsen and Mangerud (1987), Mangerud (2004) and Påsse and Andersson (2005). Ages in ^{14}C ka. Profile line indicated in Figure 1. Upper panel: *shore lines of different ages.* Middle panel: *alternative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

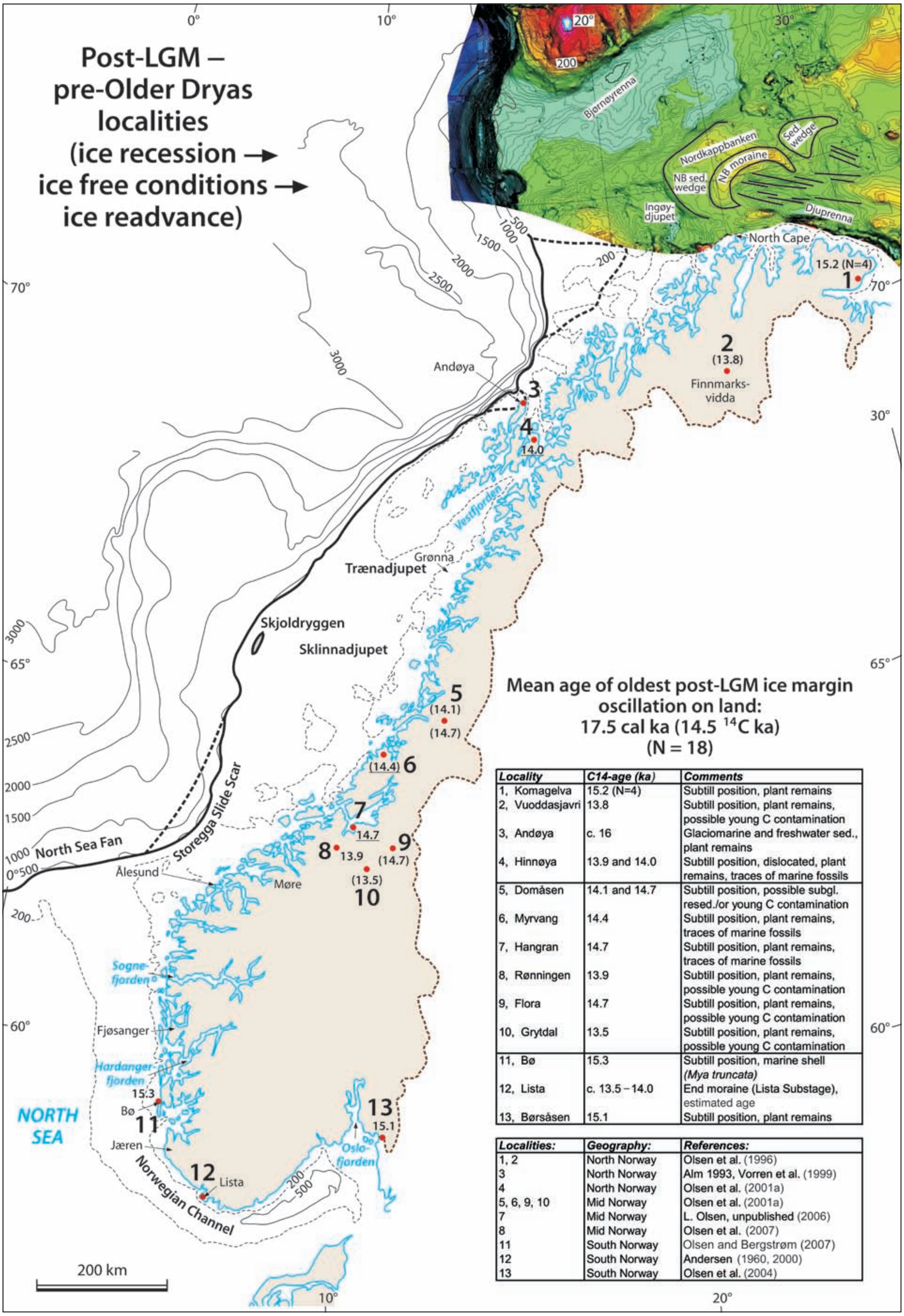
The same result with downdraw and changed geometry of the ice sheet would also be expected if the ice streams developed with initiation in the proximal parts (e.g., in the Oslofjord–Skagerrak area), where big subglacial water bodies, if they existed may have been suddenly drained and caused a considerable 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

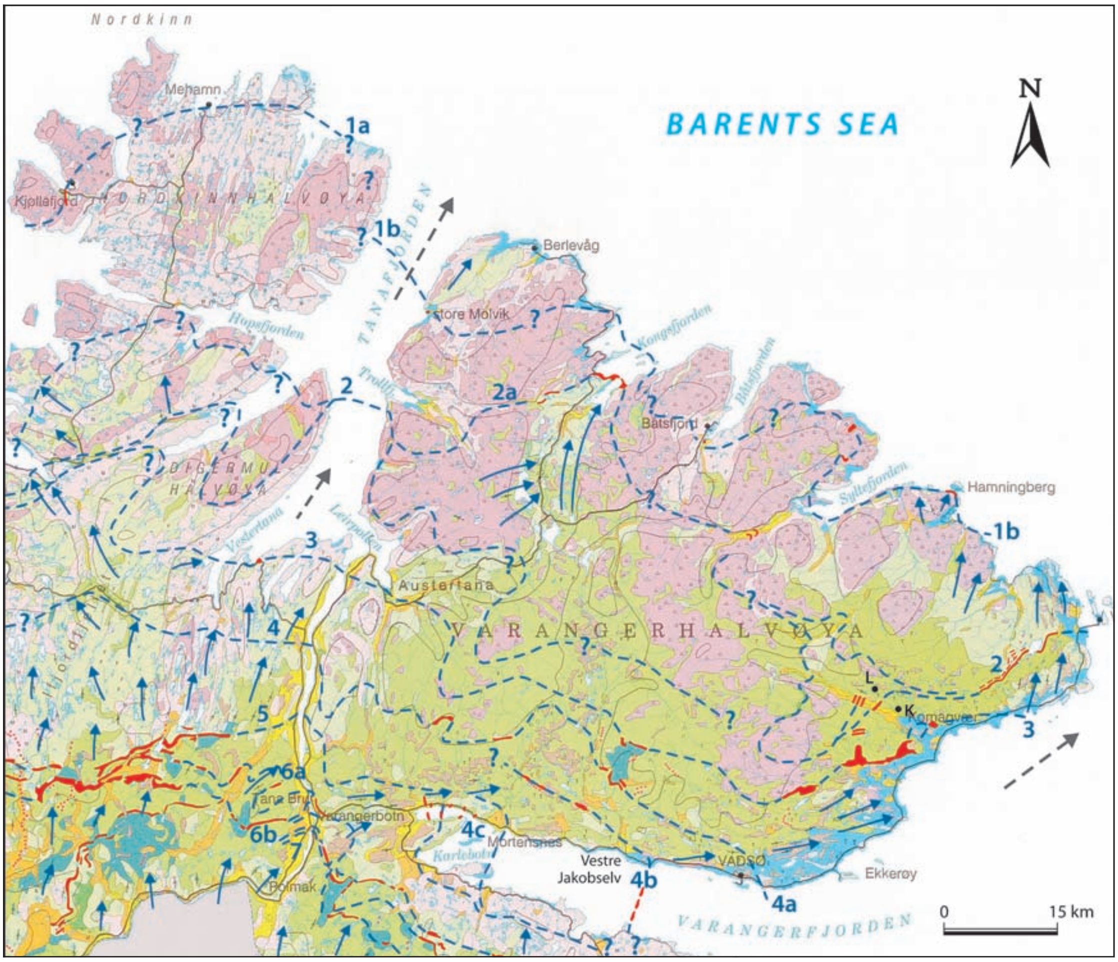


Figure 21. Inferred ice margins (1a, 1b, 2a, 2, 3, 4a, 4b, 4c, 5, 6a, 6b and 7) in northeast Finnmark, northern Norway during the Late Weichselian. Modified from Sollid et al. (1973) and Olsen et al. (1996b). 1a, 1b= The Risvik Substage, and 2, 2a= The Outer Porsanger - Vardø Substage.

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

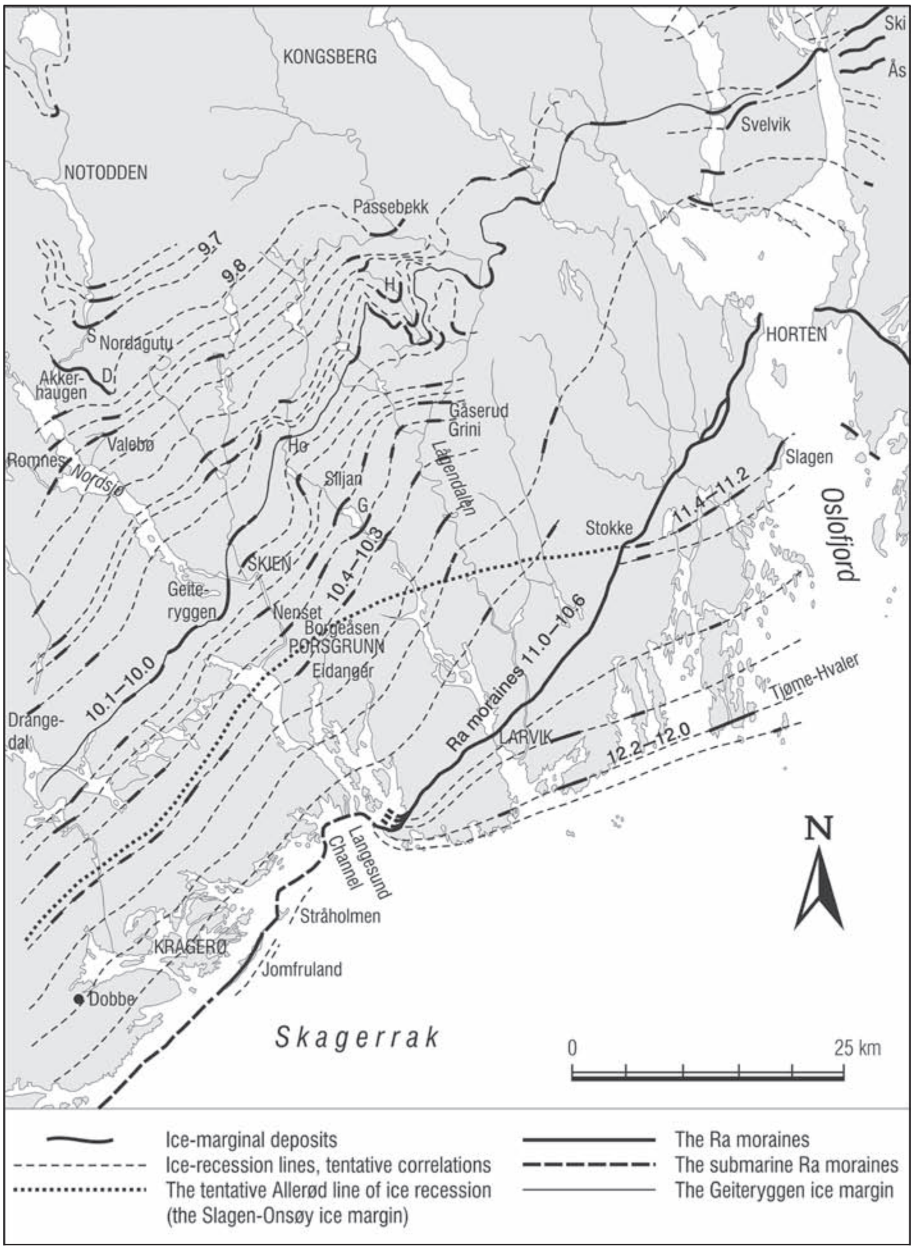
The ice retreat during the Allerød interstadial is generally assumed to have reached the heads or innermost parts of most fjords in Norway. Dates of marine shells of Allerød age and stratigraphical evidence supports this interpretation for southern Troms (Vorren and Plassen 2002, Bergstrøm and Olsen 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),

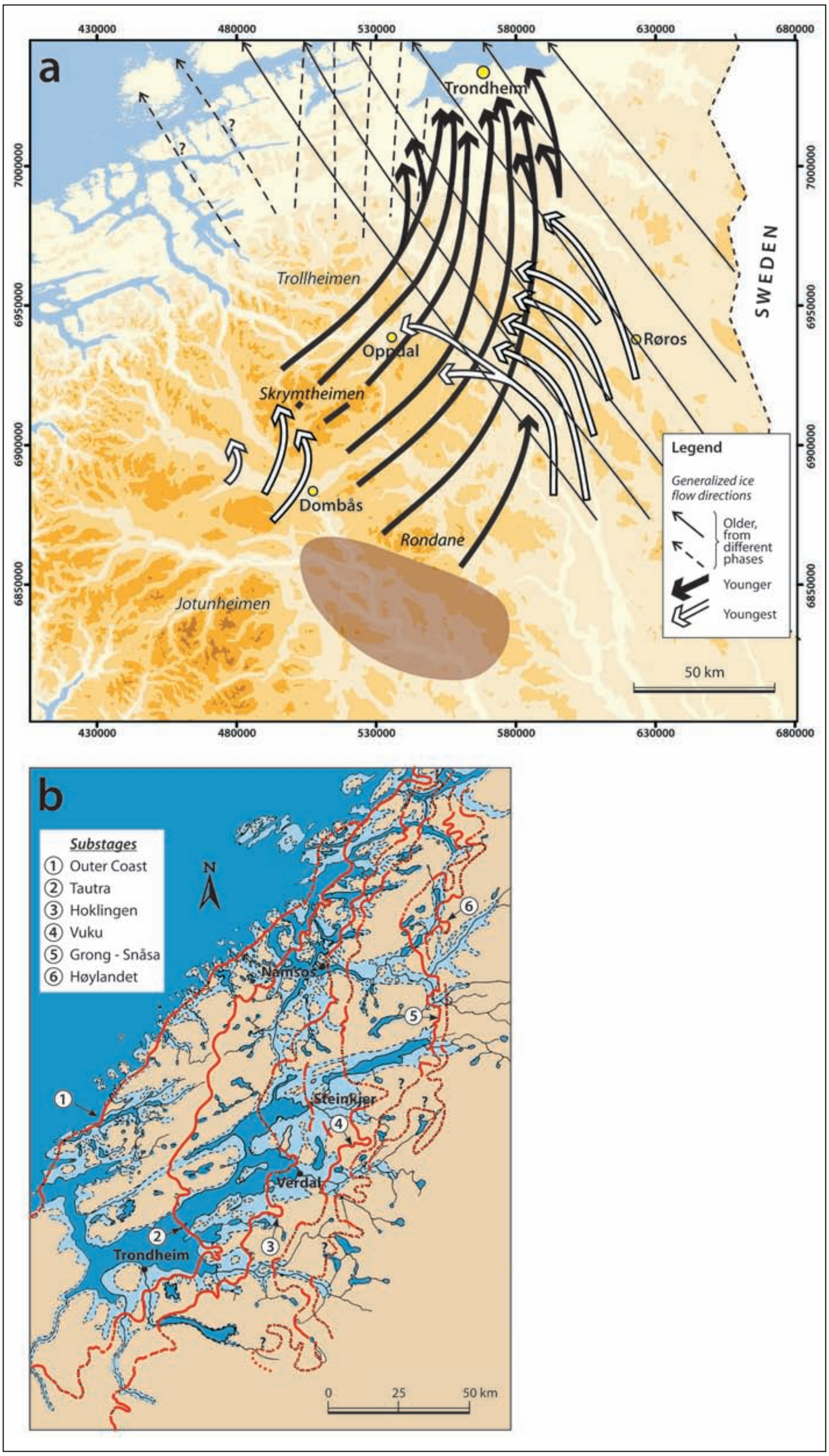


Figure 24. (a) Generalised ice-flow directions of southern central Norway during the Late Weichselian, based on glacial morphological features (modified from Follestad and Fredin 2007). The older and oldest phases represent LGM–late-glacial ice flow prior to the Allerød and Younger Dryas (Tautra Substage) (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

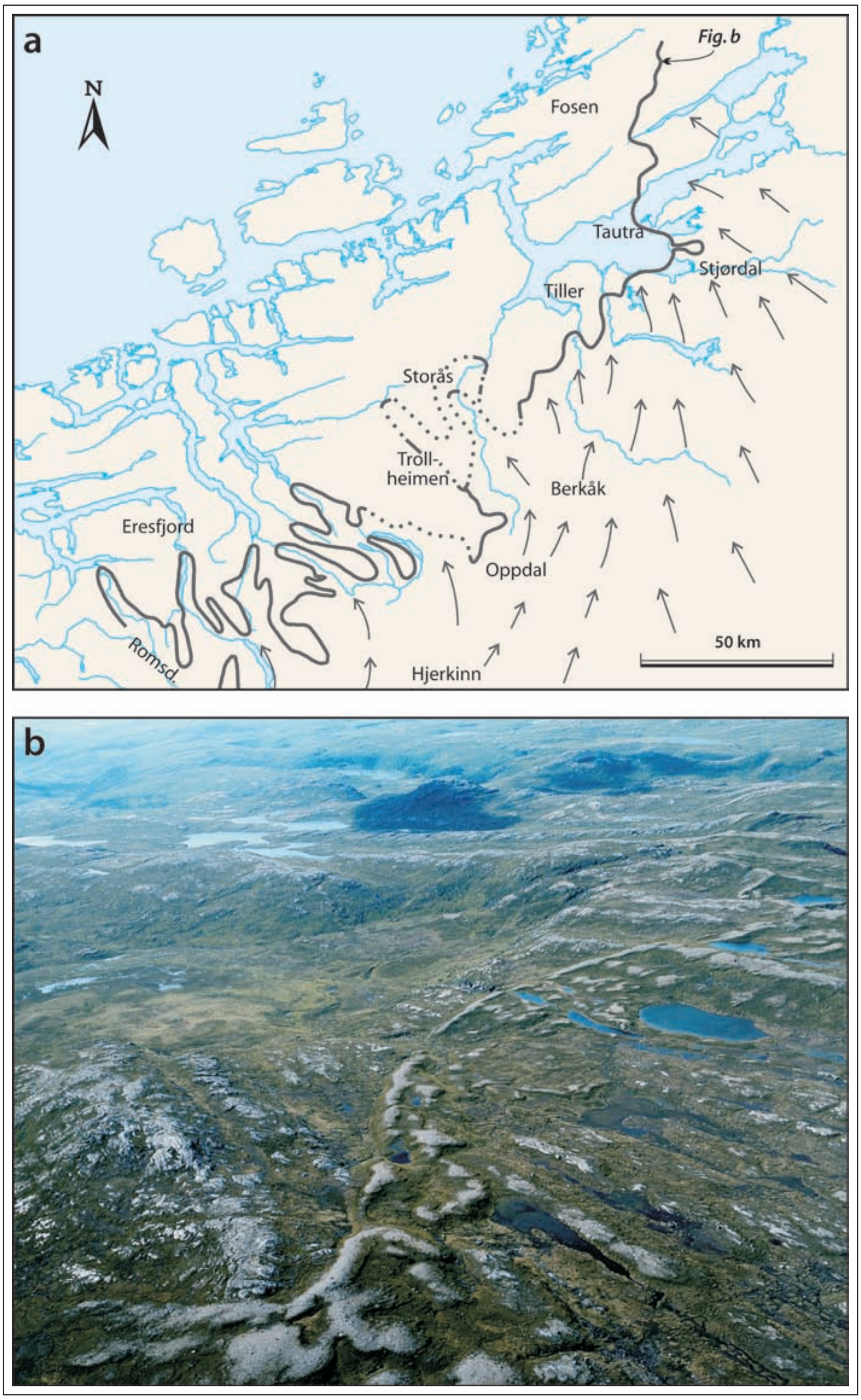


Figure 25. (a) Map of the reconstructed Younger Dryas ice margin represented by the Younger Dryas (Tautra) moraines from Trondheimsfjorden in the north to Romsdalen–Hjerkinn in the south, modified from Follestad (2005b). The stippled lines around Trollheimen indicate more uncertainty in the correlations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

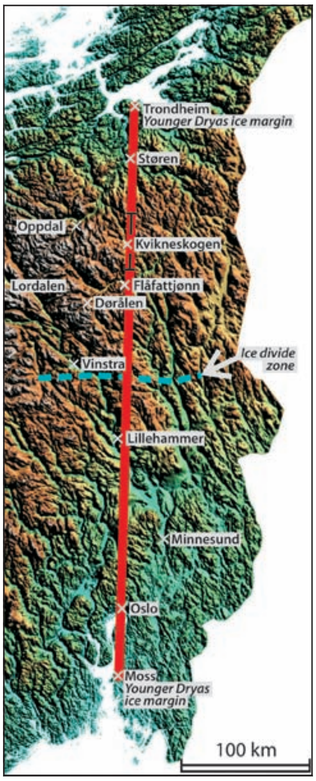


Figure 26. Ice-surface profiles for the early Younger Dryas (Tautra) and late Younger Dryas (Hoklingen) substages from Trondheim to Moss are outlined. Location of site areas and transect of profiles are shown to the left. After Olsen et al. (2007). Scenarios with and without glacial isostasy are indicated. 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).

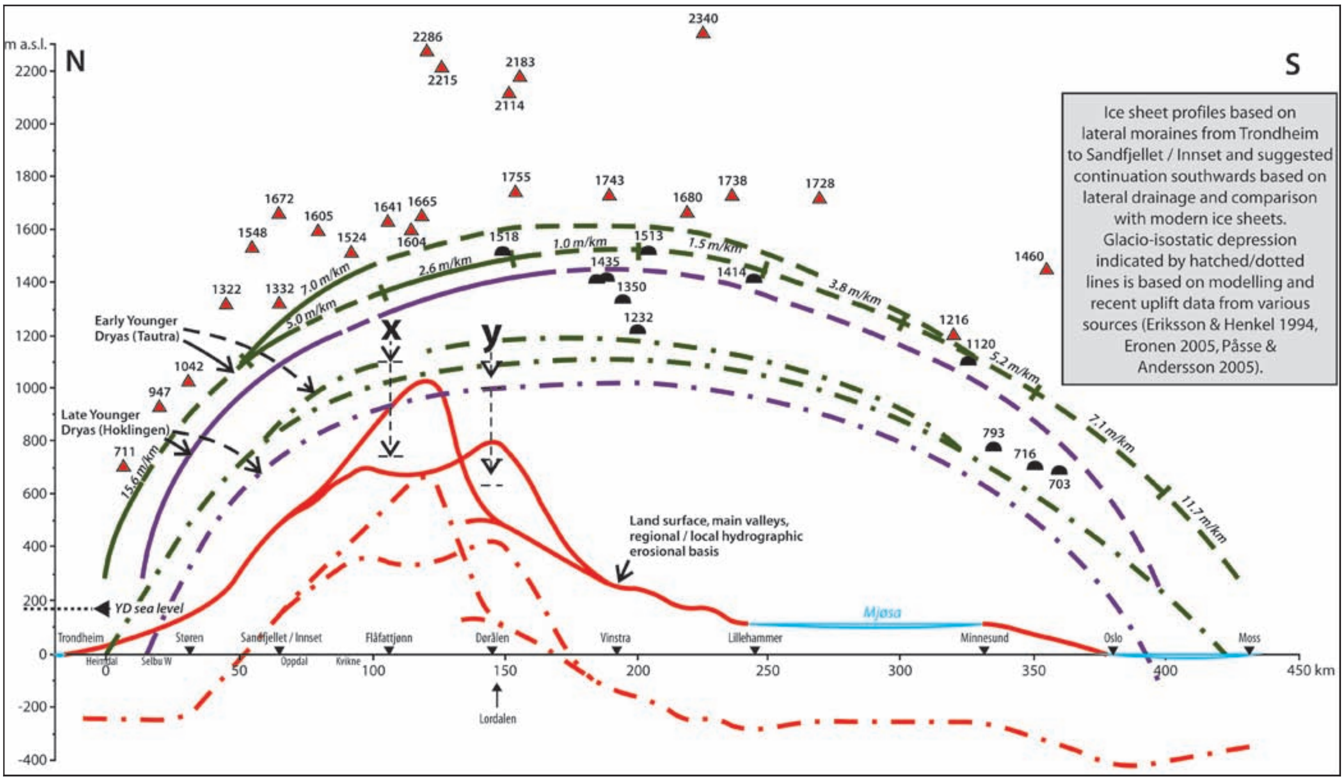


Figure 27. The distinct late Preboral Ølfjell Moraine at Saltfjellet, North Norway (Sveian et al. 1979). View from Ølfjellet towards the north. Photograph by S. Gjelle.

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).

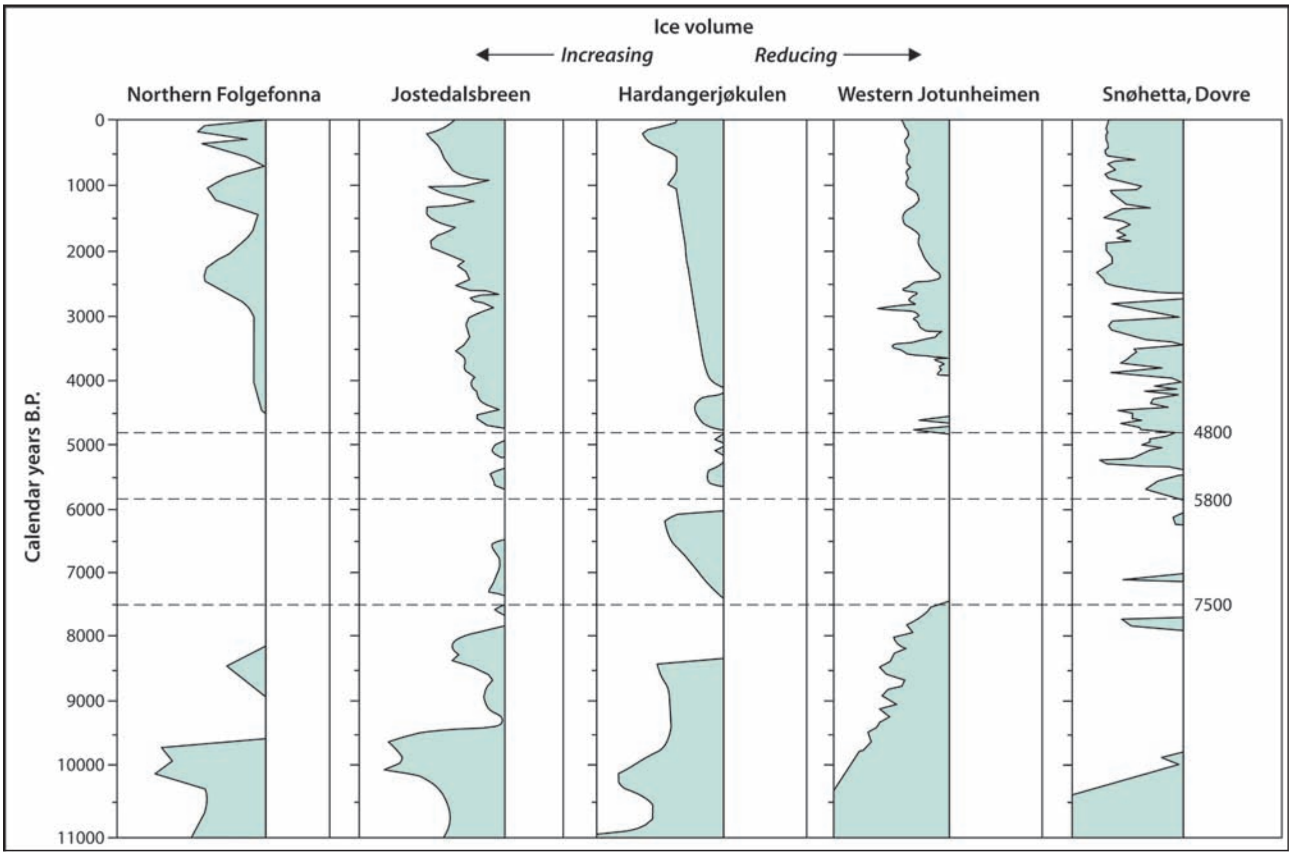


Figure 28. Variations in ice volume for five glaciers in southern Norway during the Holocene. Slightly modified from Larsen et al. (2003). All glaciers seem to have been totally deglaciated (almost no ice left) in short periods around 7500, 5800 and 4800 calendar years before present (B.P.) (stippled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

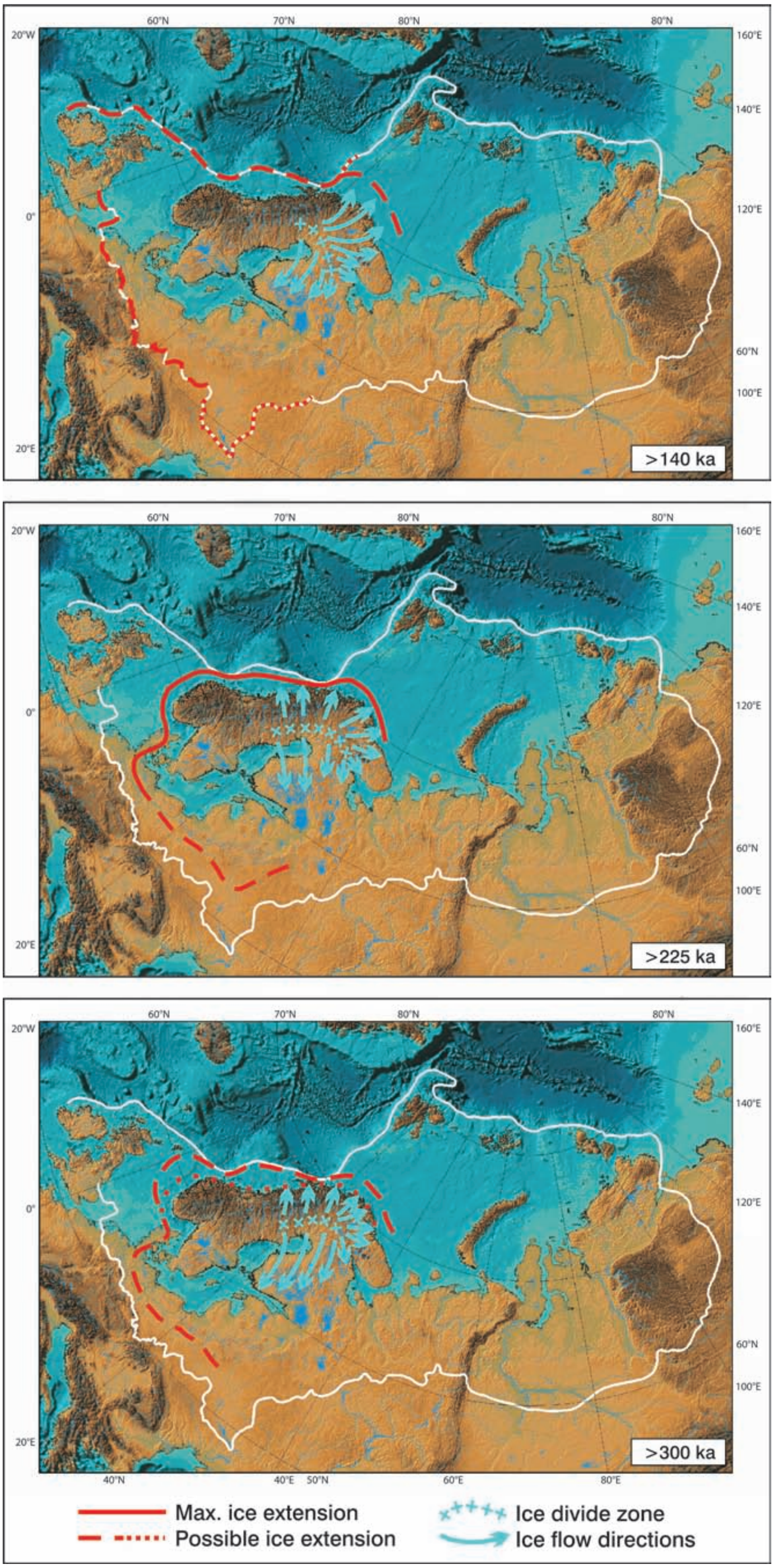


Figure A1. Glacier extension during three pre-Weichelian stages, thought to correspond to MIS 6 (upper panel), MIS 8 (middle panel) and MIS 10 (lower panel), based mainly on stratigraphical data from northern and central Fennoscandia. After Larsen et al. (2005). The correlation with data from the adjacent 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

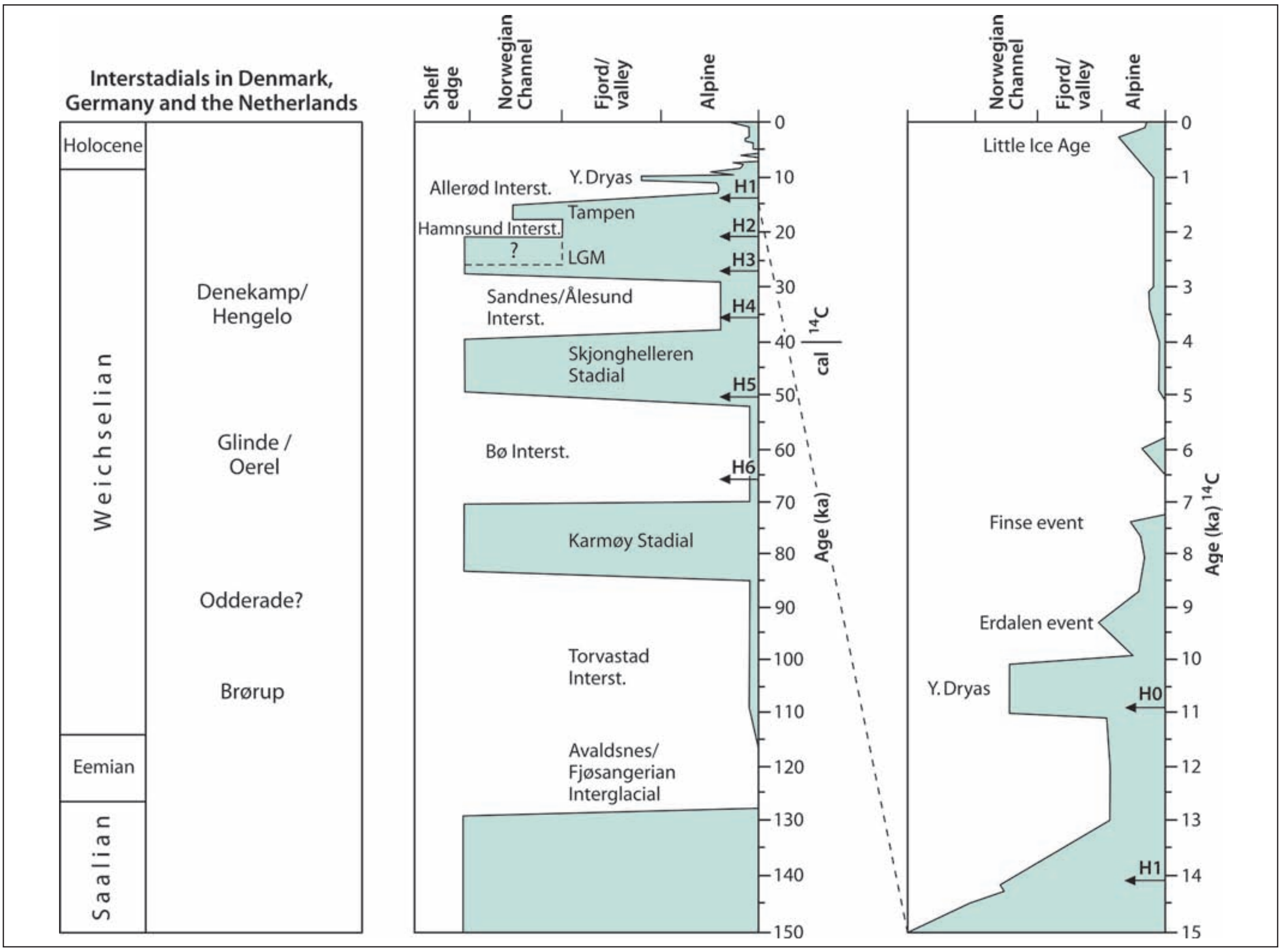


Figure B1. Glacial curves for the last 150,000 years from southwest Norway and adjacent sea-bed areas. H0, H1, H2, H3, H4, H5 and H6 indicate suggested positions of Heinrich events H0–H6. Modified from Sejrup et al. (2000). Note that the age scales are in ^{14}C ka at ages <40–45 ka and cal ka for older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

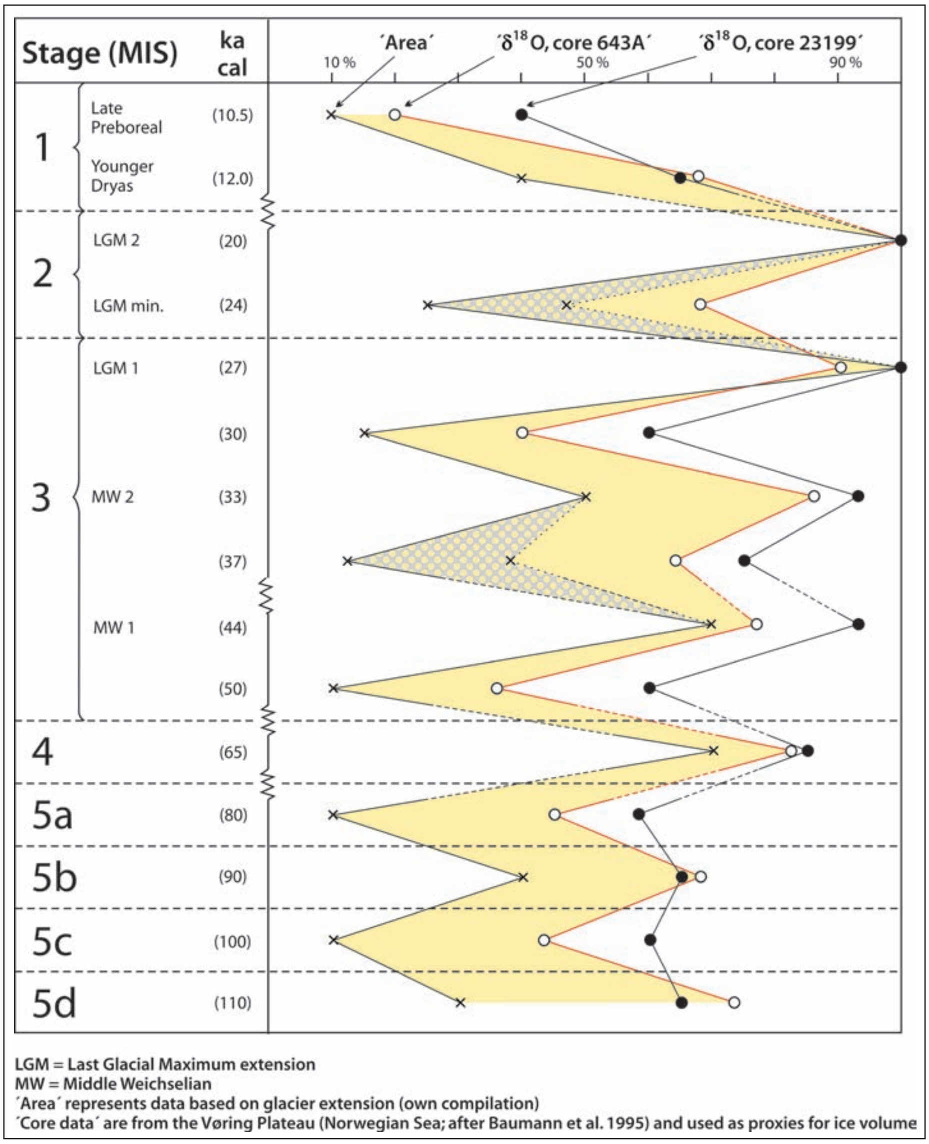


Figure B2. Relative size variations of the Fennoscandian/Scandinavian ice sheet, 110–10 ka. Comparison between horizontal ice extension and ice volume. The area glacier variations are in accordance with Figure 9 (main text), and the ice-volume data are from the Norwegian Sea (Baumann et al. 1995). After 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.

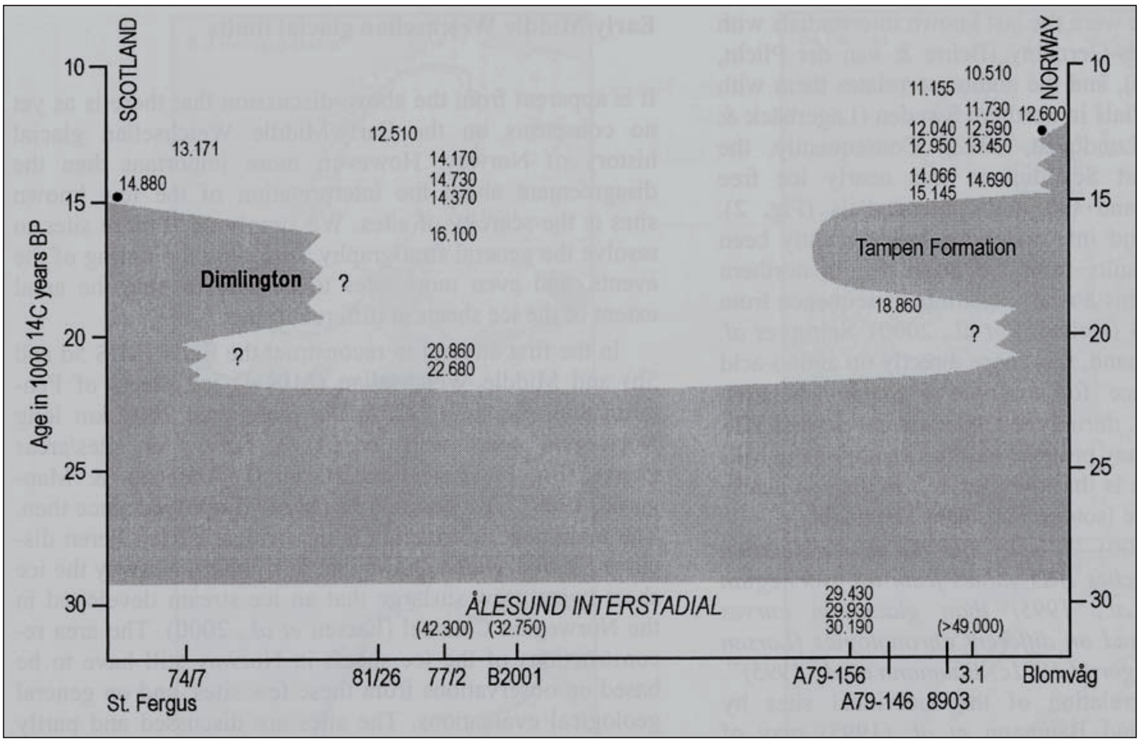


Figure B3. Glacial (time–distance) curve for the British and Scandinavian ice sheets across the North Sea during the LGM interval. The two ice sheets are supposed to have met before c. 22.7 ^{14}C ka (c. 26.7 cal ka). Ages are in ^{14}C yr BP. After Sejrup et al. (2000).

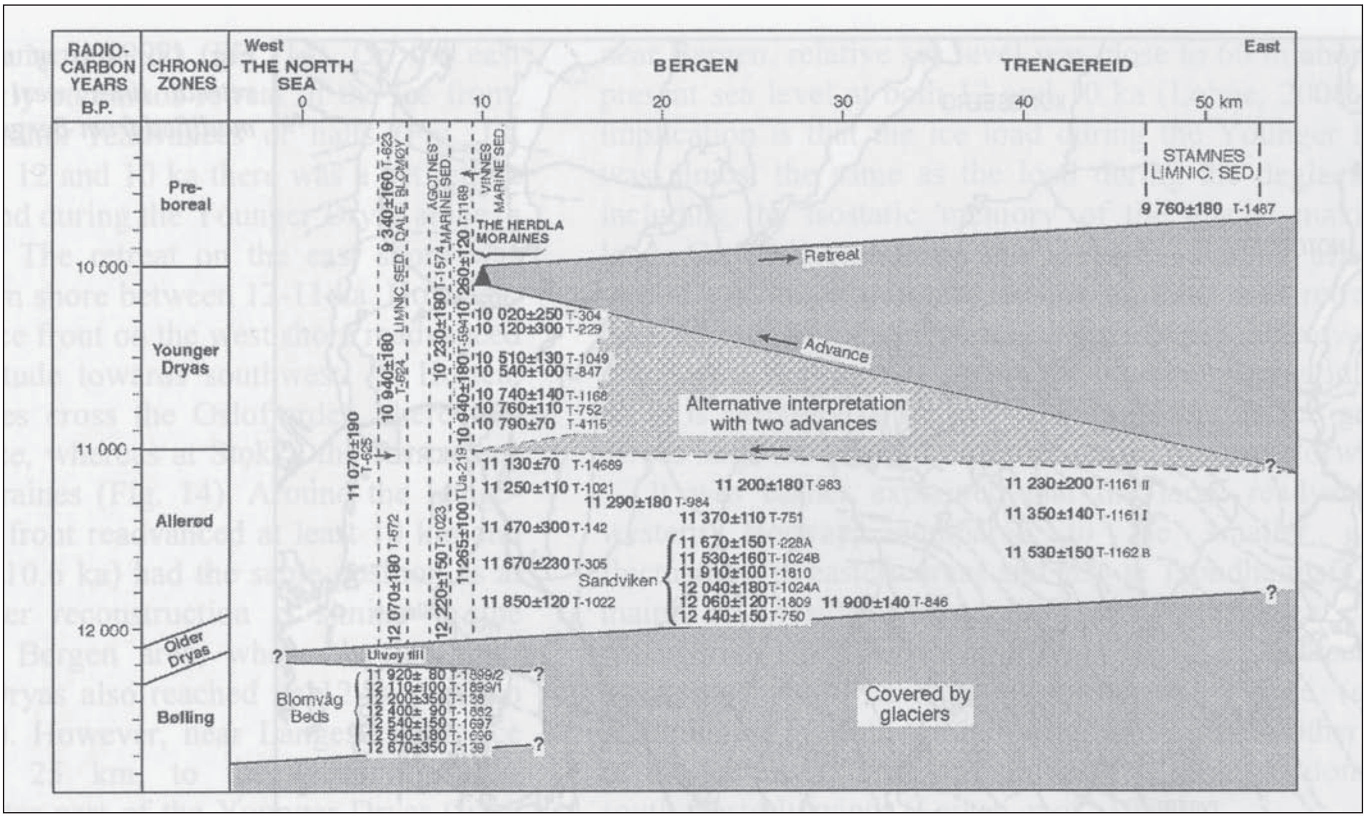
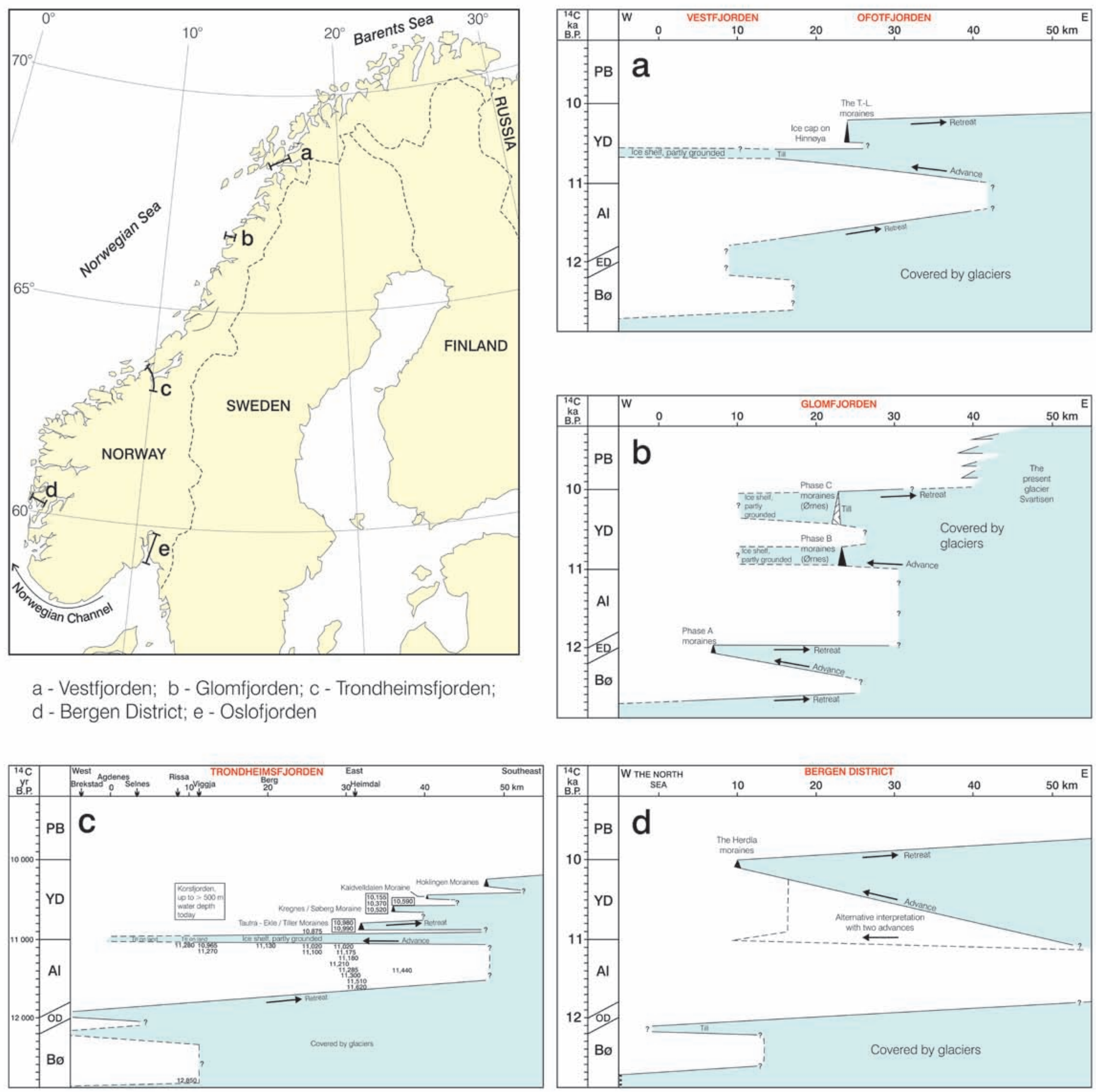


Figure B4. Ice-front fluctuations in the Bergen district, from Mangerud (2000). Ages are in ^{14}C yr BP. A generalised version of this diagram is included as diagram d in Figure B5.



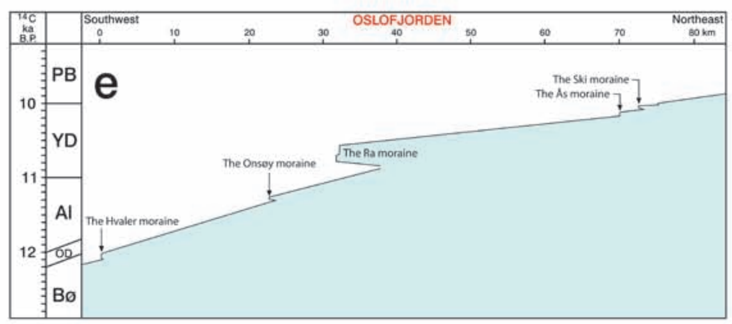


Figure B5. Ice-front fluctuations in various parts of Norway. a–Vestfjorden, b–Glomfjorden, c–Trondheimsfjorden, d–Bergen district, and e–Oslofjorden, eastern shore. Modified from Olsen et al. (2001d, 2007). A photograph of a cirque moraine formed during phase C as indicated in diagram b is shown in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

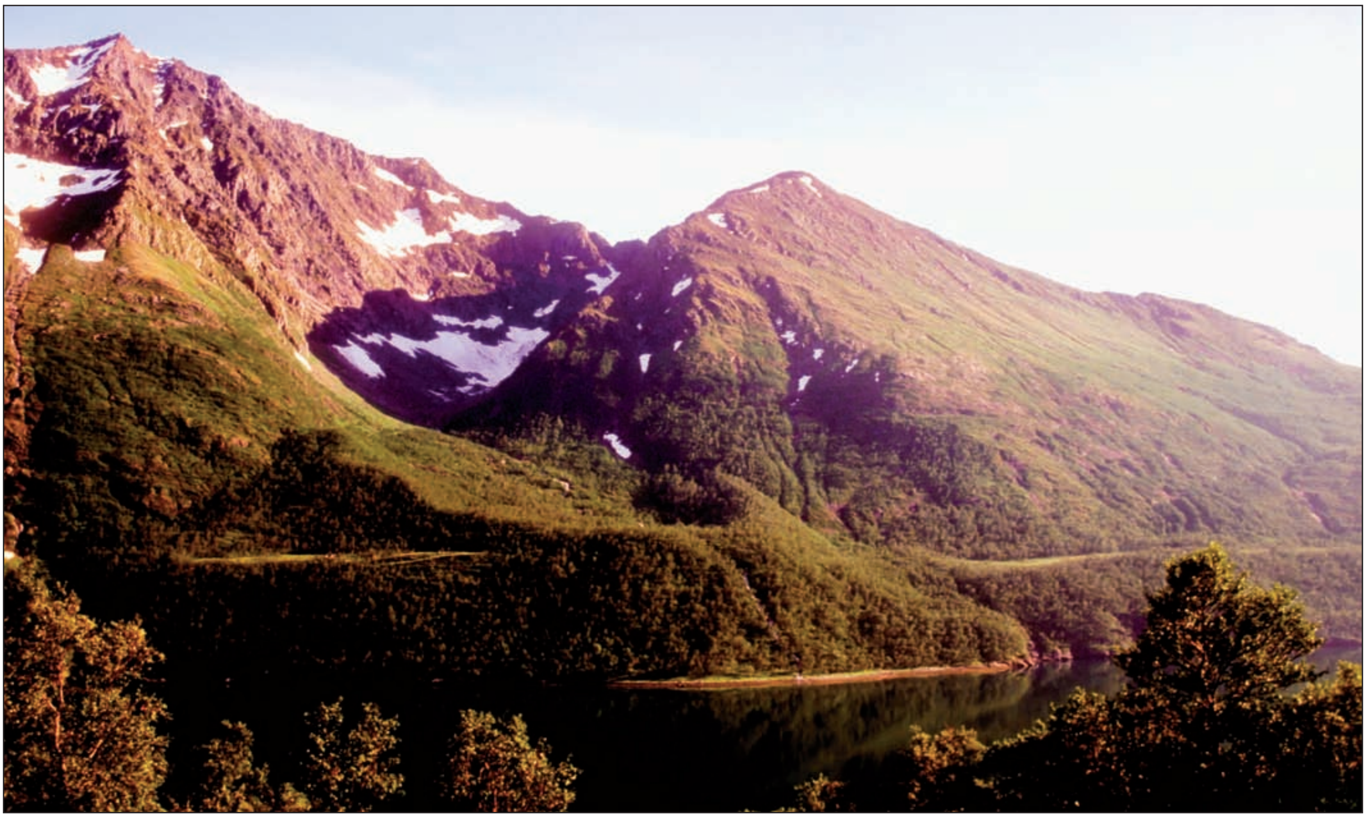


Figure B6. Cirque (-glacier) moraine at the southern side of Holandsfjorden west of Svartisen in Nordland (see Figure 3, the main text). The cirque moraine from late Younger Dryas (during phase C in Figure B5, diagram b) is located partly on a shoreline (c. 100 m a.s.l.), which was eroded in bedrock 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).