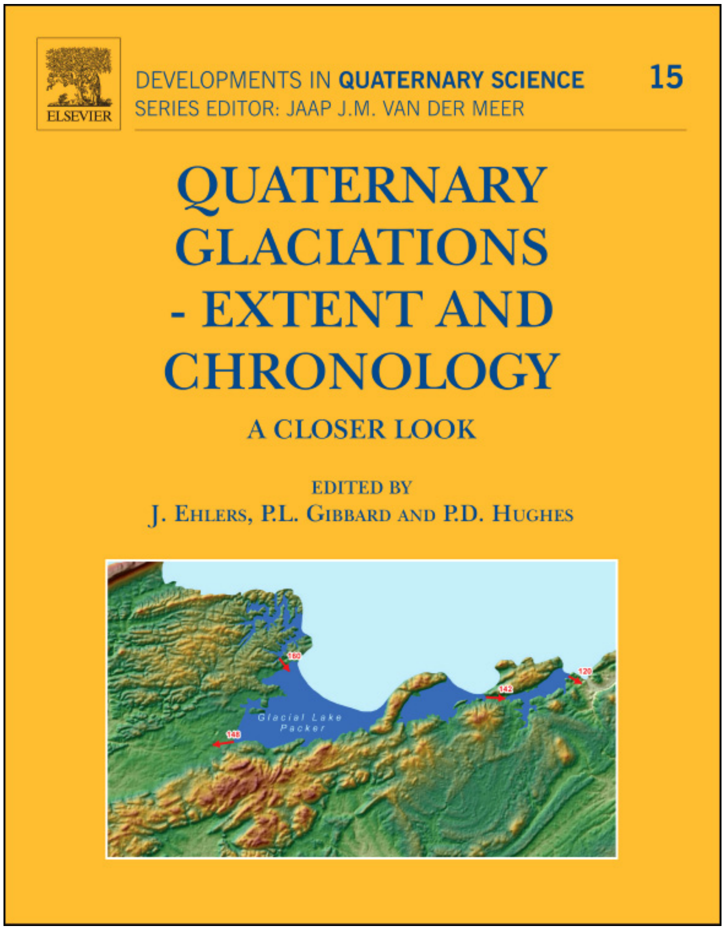

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Chapter 22

Glacial History of Norway

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22.1. INTRODUCTION

This review is in principle an updated version of a similar chapter in the previous book edition (Mangerud, 2004). This time, we have to a large extent discussed other aspects of the glaciation history and most of the discussions and references to older papers that were included in Mangerud (2004) have been omitted.

Glaciers in the Norwegian and Swedish mountains certainly have acted as nuclei for Scandinavian Ice Sheet growth several tens of times during the Quaternary, and probably even during the Pliocene. The ice sheets, and during milder periods also smaller glaciers, have had a profound impact on the Norwegian landscape. The many deep fjords, long U-shaped valleys, cirques and thousands of lakes in overdeepened bedrock basins are the results of glacial activity. The vertical linear glacial erosion along several fjords amounts to 1500–2000 m, whereas it probably is considerably less than 100 m on some plateaux and summits; a description and discussion of the pattern of glacial erosion is given by Kleman et al. (2008). A mean bedrock lowering of ~520 m through the Quaternary glaciations has been calculated for a large area of Mid-Norway (Dowdeswell et al., 2010). However, the bedrock landforms cannot by themselves be used to unravel the full glaciation history. On the contrary, the glaciation history inferred from other data is used as one of the main factors explaining the observed pattern and inferred rates of glacial erosion.

22.2. EARLY AND MIDDLE QUATERNARY GLACIATIONS

The last ice sheets that covered the Norwegian land masses during the last (Weichselian) ice age have removed most of the older Quaternary deposits on land. Accordingly, pre-Weichselian glacial deposits are found only in a few sites (Mangerud, 2004). The older glacial history has to be deciphered from sea floor records in the deep Norwegian Sea and from deposits along the continental margin and from the North Sea region with the adjacent land masses to the south.

Ice-rafted debris (IRD) dropped from "Norwegian" icebergs into the deep sea beyond the coast represents the most continuous records of former glaciations. However, the ice sheets may have reached a considerable size over Norway–Sweden before they reached the sea and released icebergs. One should also keep in mind that the distribution and amount of IRD are also influenced by other factors, such as ocean currents and sea surface temperatures. IRD in the records from the Nordic Seas may also stem from Greenland or from sea ice. Kleiven et al. (2002) presented a synthesis of IRD results from several cores covering the period 3.5–2.4 Ma ago, and they concluded that there had been marked expansions of the Greenland Ice Sheet at 3.3 Ma and of the Scandinavian Ice Sheet from 2.75 Ma. Cores from the Vøring Plateau (Fig. 22.1) outside the Norwegian coast also contain some minor IRD pulses starting as early about 12.6 Ma, and another at about 7.2 Ma (Fronval and Jansen, 1996).

There are several recent reviews and syntheses of the glaciation history of the Norwegian continental shelf (Dahlgren et al., 2005; Hjelstuen et al., 2005; Nygård et al., 2005; Rise et al., 2005; Sejrup et al., 2005; Ottesen et al., 2009). The architecture of the deposits is somewhat different on the North Sea Fan at the mouth of the Norwegian Channel (Fig. 22.6) as compared with the architecture in areas to the north and south of the fan. The Naust Formation is the youngest formation on the Mid-Norwegian continental shelf in the area between 64 and 68°N, that is, north of the North Sea Fan. It consists of a series of prograding wedges (Fig. 22.2) where the base is considered to have an age of ca. 2.8 Ma (Ottesen et al., 2009), that is, close to the recorded major increase in IRD in the Norwegian Sea and the onset of the Quaternary according to the new definition of the boundary (Gibbard et al., 2009). However, there are still significant uncertainties concerning the exact age of the lower boundary of the Naust Formation as well as the internal subunits. The Naust Formation is more than a

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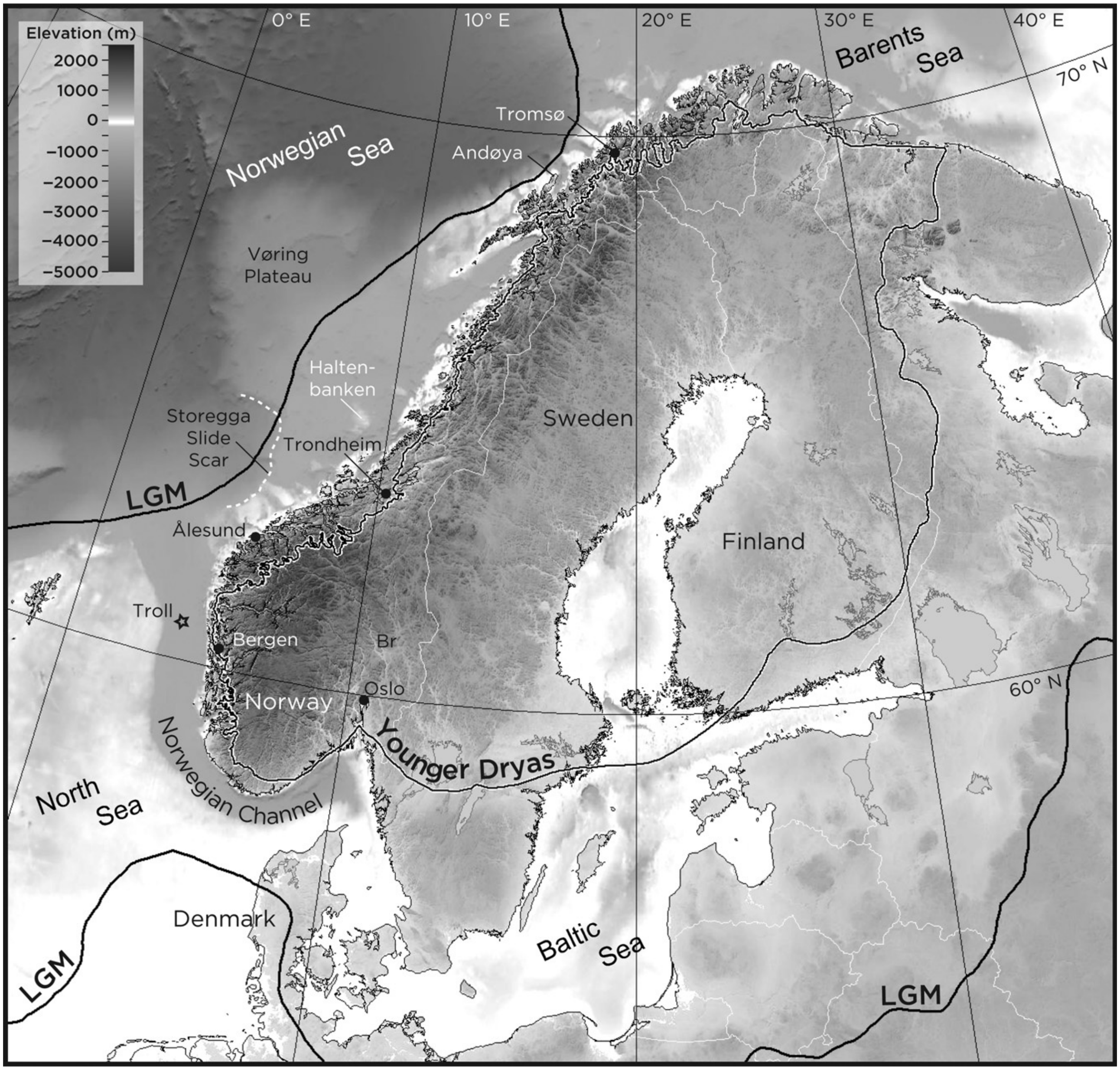

FIGURE 22.1 Map of Scandinavia with adjacent lands and seas. Note that the main mountain range starts in central south Norway and continues northeastwards along the Norwegian–Swedish border. The (asynchronous) ice limit at the Last Glacial Maximum (LGM) (Svendsen et al., 2004) and also the Younger Dryas moraines (Gyllencreutz et al., 2007) are marked. Br is the Brumunddal site. Digital elevation model is based on the SRTM30PLUS-dataset (Becker et al., 2009).

kilometre thick over a large area, and the shelf edge prograded up to 150 km during accumulation of this formation (Dahlgren et al., 2005; Rise et al., 2005; Ottesen et al., 2009). Independent of the remaining chronological uncertainties it is clear that the Naust Formation represents a period with more rapid deposition as compared to the underlying units. Rise et al. (2005) identified till in the oldest part of the Naust in boreholes on Haltenbanken (Fig. 22.1), off Mid-Norway, and Ottesen et al. (2009) described about 1.5 Ma old glacial flutes crossing the shelf. They conclude that large parts of the thick sediment wedge that has been dated to 2.8–1.7 Ma were formed during several ice sheet advances to the palaeo-shelf edge. Contrary to this view, several other investigators (Jansen et al., 2000; Dahlgren et al., 2005; Hjelstuen et al., 2005; Sejrup et al., 2005) consider that this older part of the Naust reflects some smaller mountain-centred ice sheets and that the ice margin reached the shelf edge for the first time about 1.1 Ma. However, there is a full agreement that the Naust Formation reflects a number of glaciations that at least covered a large part

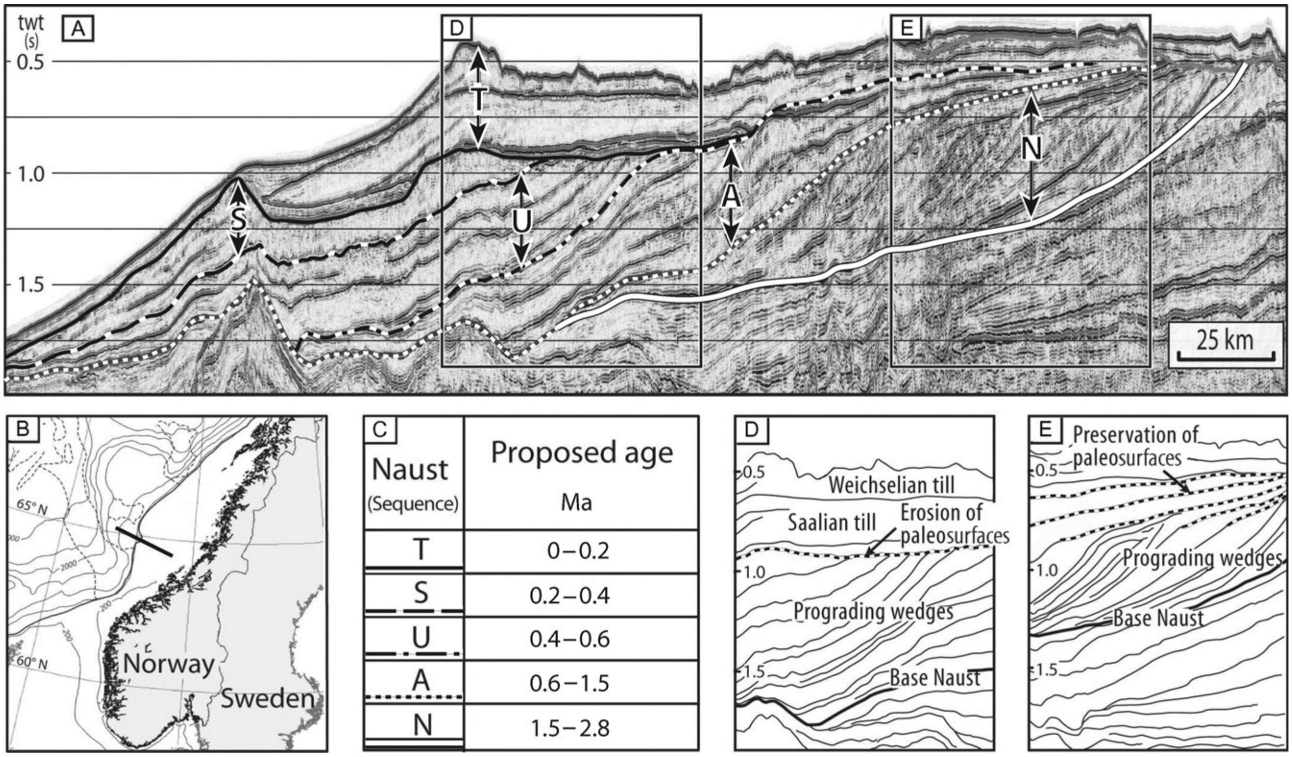

FIGURE 22.2 (A) Seismic-stratigraphical section across the mid-Norwegian continental shelf and slope. The Naust Formation (marked with letters N, A, U, S and T for subunits) is the glacigenic influenced wedge. Twt, two-way travel time; 1 s equals ~1000 m of sediment thickness. (B) Location of the profile. (C) Subdivisions and proposed ages (Ottesen et al., 2009) for the Naust Formation. The boundaries are marked with different line types used in (A). (D) Interpreted part of the section showing truncation of prograding wedges and palaeo-shelf surfaces. (E) Interpreted part showing preservation of wedges and palaeo-shelf surfaces. Stippled lines, see (D) above. Modified from Dowdeswell et al. (2007).

of mainland Norway. On the European continent further to the south, only one or two glaciations older than 0.5 Ma have been proven (Ehlers et al., 2004; Houmark-Nielsen, 2004; Laban and van der Meer, 2004). During the period 2.7–1.1 Ma ago that was dominated by the 41 ka glaciation cycle, the Scandinavian mountains were repeatedly hosting sizeable ice sheets that eroded Norwegian fjords (Mangerud et al., 1996). Only during the past million years or so, when the 100 ka cycles dominated did the Scandinavian Ice Sheet for short periods expand far south of Scandinavia.

The oldest identified till in the Norwegian Channel is the Fedje Till that is dated to about 1.1. Ma (Sejrup et al., 1995, 2005). In the period from 1.1 Ma to marine isotope stage (MIS) 12 (about 500 ka), there were more limited glaciations. Starting with MIS 12, the North Sea Fan became the main depo-centre of glacial sediments from southern Scandinavia (Nygård et al., 2005; Sejrup et al., 2005). Four pre-Weichselian glacigenic debris-flow packages that were correlated with MIS 12, 10, 8 and 6, respectively, are identified in the fan (Hjelstuen et al., 2005; Nygård et al., 2005; Sejrup et al., 2005), indicating that the ice stream each time reached the mouth of the channel. There is a general agreement that the ice sheet reached the shelf edge further north on the continental shelf several times during the past 0.5 Ma (Fig. 22.3) also. This includes the Elsterian, Saalian and other glaciations that expanded far south on the continent (Dahlgren et al., 2005; Hjelstuen et al., 2005; Rise et al., 2005; Ottesen et al., 2009). It is interesting to note that Dowdeswell et al. (2006) identified a different and longer ice stream path on the shelf off Mid-Norway during the Saalian than during the Late Weichselian, which also implies thicker ice up-flow.

The extension of the Scandinavian Ice Sheet towards west was probably several times limited by the water depth and thus calving and not climate because grounded ice sheets cannot extend into deep open water. Therefore, the 150-km progradation of the Mid-Norwegian shelf during deposition of the Naust Formation made possible a further ice sheet extension during the younger than during the earlier glaciations, also favouring thicker ice sheets over the main land.

22.3. THE LATE QUATERNARY GLACIATIONS—THE WEICHSELIAN

By definition, the Weichselian starts at the end of the Eemian, corresponding approximately with the end of the MIS 5e (or early MIS 5d). The glacial history during this

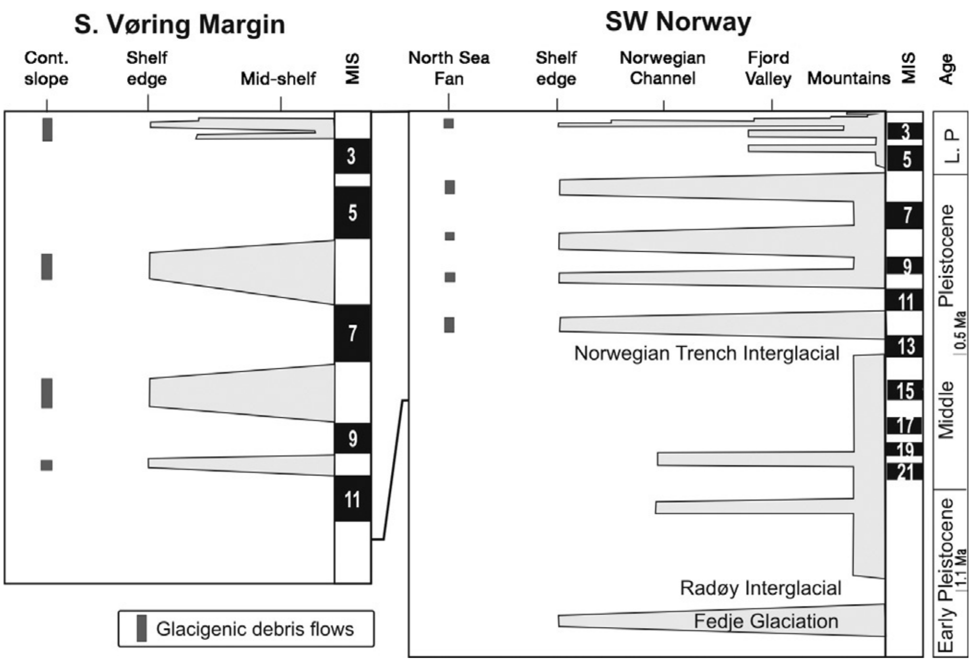

FIGURE 22.3 Schematic time–distance diagrams for the margin of the Scandinavian Ice Sheet through the Quaternary. The left diagram (South Vøring, Fig. 22.1) shows the Mid-Norwegian Shelf, and the right diagram shows SW Norway from the mountains through the Norwegian Channel to the shelf edge. Slightly modified from Hjelstuen et al. (2005).

period is for obvious reasons better known than for older periods. However, considering that the Late Weichselian (Last Glacial Maximum (LGM)) ice sheet advance removed most pre-existing sediments, our knowledge of the foregoing Weichselian remains fragmentary. The interpretation of the timing and extent of these older glaciations are based on observations from some few localities. The available data are insufficient to allow accurate mapping of the ice sheet limits, but they nevertheless give a rough idea of the ice sheet dimensions (Fig. 22.5). We have chosen to subdivide this chapter into two main sections: in the first one, we describe the Early and Middle Weichselian, and in the second, we describe the ice sheet extend during the LGM and the last deglaciation.

22.4. THE EARLY AND MIDDLE WEICHSELIAN

22.4.1. The Period of MIS 5e to 4

Dating events beyond than the range of the radiocarbon method are problematic. Different scientists have therefore disagreed on the age of some of these deposits and thus also the correlation between different sites. We refer to the discussion in Mangerud (2004); here, we will mainly describe some key sites and review some new results. Figure 22.4 shows our favoured glaciation curve for south-western Norway. The oldest part of the curve is based mainly on the stratigraphy at the key site Fjøsanger (Mangerud et al., 1981). The correlation of the Fjøsangerian Interglacial with the Eemian is well established. Conformably, overlying the interglacial strata is a thick glaciomarine silt unit (Gulstein in Fig. 22.4), indicating that a large glacier ended somewhere in the fjord close to the Fjøsanger site. The glaciomarine silt is covered by a non-glacial beach deposit (Fana interstadial) followed by a thick till bed, the Bønes Till (Fig. 22.4). The till contains a large number of sediment clasts, molluscs and wood remains that must have been picked up from the 80-m-deep fjord basin adjacent to the site. The least certain part of this chronology is the correlation of the Fana interstadial with the Brørup interstadial (MIS 5c). If our correlation should be wrong, we find it most probable that Fana is older than MIS 5c and that the Bønes Till was deposited during MIS 5d, as originally proposed by Mangerud et al. (1981).

Another key site is located in Brumunddal in central parts of southern Norway (Fig. 22.1). At this site, a well-defined layer of peat is interbedded between two till beds (Helle et al., 1981). The pollen stratigraphy of the peat shows a succession starting from open pioneer vegetation to shrubs with some trees including *Larix*, and a reversion to arctic tundra near the top of the peat, that is, a cold–mild–cold climate cycle. The correlation of the peat with the Brørup seems plausible (Helle et al., 1981), although not unambiguously proven. If correct, the site provides evidence for the existence of a pre-Brørup till that most likely postdates the last interglacial and thus suggests a MIS 5d ice sheet advance across the site (Fig. 22.4).

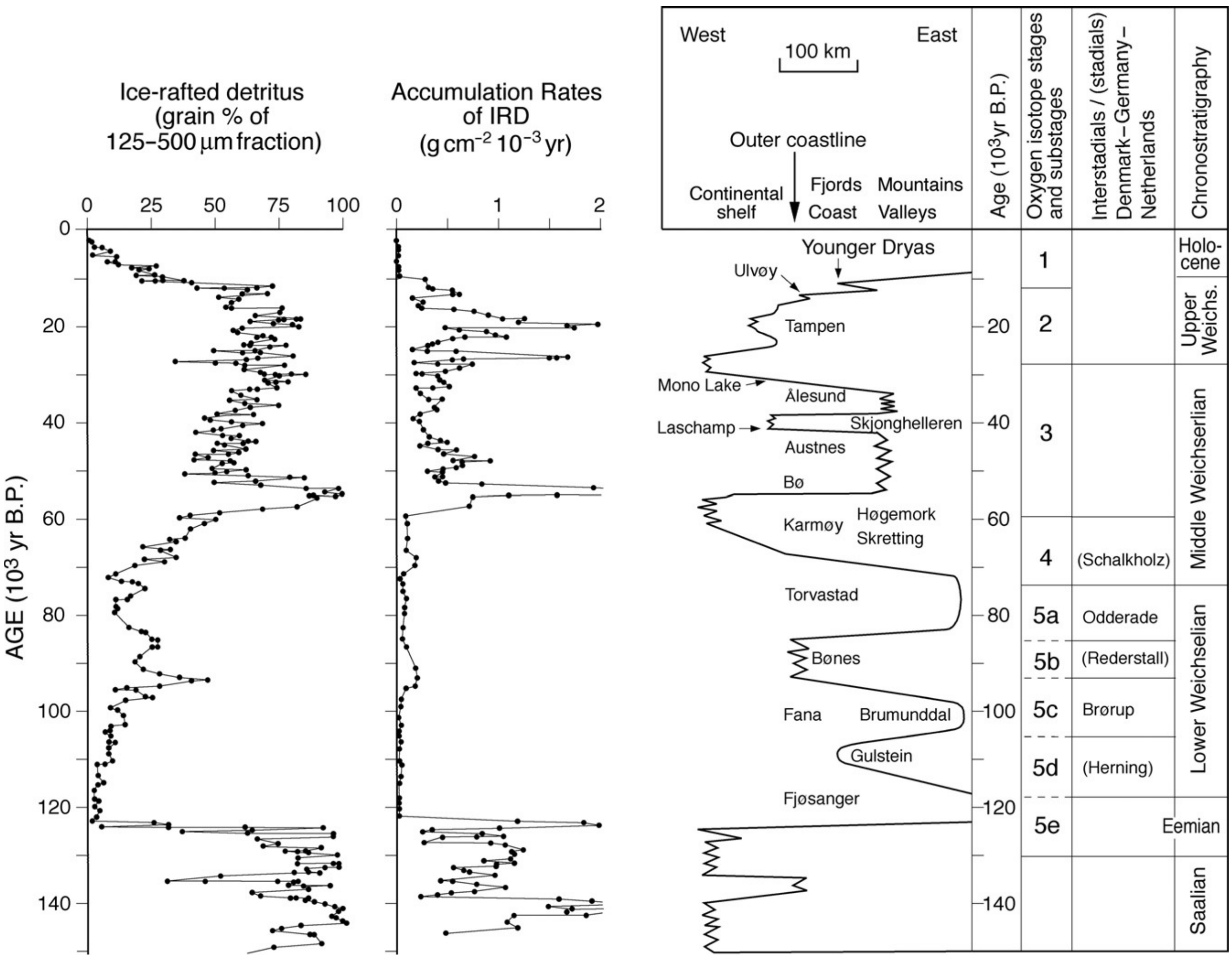

FIGURE 22.4 The time–distance diagram to the right is a schematic glaciation curve for the south-western flank of the Scandinavian Ice Sheet through the Weichselian, modified from Mangerud (2004). Names on the curve represent geological sites. The Laschamp and Mono Lake palaeomagnetic excursions are marked with arrows. The curve to the left is a stacked curved for grain % of IRD in five cores, and the middle curve a stacked record of accumulation rates in three cores; all cores collected from the Vøring Plateau (Fig. 22.1; Baumann et al., 1995). The timescale is (in principle) in calendar years, but the curves are partly dated and correlated with different methods.

The MIS 5e–MIS 4 part of the glaciation curve (Fig. 22.4) is recently supported by peaks in the ice-rafted detritus (IRD) record in a core collected from the Vøring Plateau (Fig. 22.1; Brendryen et al., 2010), in addition to earlier IRD curves shown in Fig. 22.4 (Baumann et al., 1995). Even though IRD is not a simple monitor of ice sheet size, iceberg rafting demonstrates that glaciers at this time extended down to sea level and probably even reached the open sea (i.e. beyond the mouth of the fjords). Brendryen et al. (2010) even correlate some of their IRD peaks with individual Dansgaard–Oeschger events; such a detailed correlation is not yet possible with the more poorly dated land record. We note that the δ^{18}O in precipitation on Greenland respond on a seasonal scale to weather and climate changes, and also some marine organisms and chemical parameters respond almost that fast. Ice sheets, and especially the build-up of ice sheets, will have a much slower response, and it will take hundreds and probably even thousands of years to build up a large ice sheet covering Norway.

Some of the most important new evidence for the ice sheet development in Norway during MIS 4 and early MIS 3 come from Denmark (Larsen et al., 2009a,b; Houmark-Nielsen, 2010). Based on several occurrences of a till with Fennoscandian erratics, Larsen et al. (2009a, b) describe the Sundsøre glacial advance from Norway which reached almost half-way down Denmark and was dated to 65–60 ka. They correlate this advance with the Karmøy advance in Norway (Mangerud, 2004; Fig. 22.4). If correct, the implication is that all of southern Norway with the adjacent shelves was ice covered during MIS 4, as indeed reconstructed by Carr et al. (2006). We also note that there is evidence to suggest that also the Barents–Kara Ice Sheet expanded at this time reaching the edge of the continental shelf west of Svalbard (Mangerud et al., 1998). Although there at present do not exist observations to test our hypothesis, we find it reasonable to assume that the Scandinavian Ice Sheet reached close to the shelf edge along the entire Norwegian coast during MIS 4 (Fig. 22.5).

Larsen et al. (2009a) demonstrate stratigraphically that the Sundsøre ice margin retreated and lacustrine and marine sediments were deposited in northern Denmark before there was another (Ristinge) ice advance across much of the country. However, the Ristinge advance came from the Baltic Sea, and it is dated to about 50 ka (Houmark-Nielsen, 2010), that is, early MIS 3. Apparently, the main dome of the ice sheet had moved east towards the Baltic Sea from its more westerly position when the Sundsøre advance occurred. There is no site in Norway that can resolve these two ice advances, and in the schematic map for MIS 4 in Fig. 22.5, we have combined the maximum limits for the two, although we acknowledge that they are asynchronous. Svendsen et al. (2004) correlated this Ristinge advance with the MIS 4 glacial maximum in the Barents Sea–Kara Sea Northern Russia area. In the entire Russian sector, the MIS 4 ice sheet was considerably larger than the LGM ice extent. The conclusion now is that the Scandinavian Ice Sheet probably was as (or almost as) extensive west and south of Norway during MIS 4 (and/or early MIS 3) as during MIS 2 (LGM).

22.4.2. Marine Isotope Stage 3

The start of MIS 3 is beyond the reach of radiocarbon dating, but ^{14}C ages are getting more and more trustworthy towards the end of MIS 3. In this section, we will not follow a chronological description, but start with the best dated part in the middle and late MIS 3, and then discuss the older part.

The coastal caves Skjonghelleren (Larsen et al., 1987), Hamnsundhelleren (Valen et al., 1996) and Olahola (Valen et al., 1995) located near Ålesund (Fig. 22.1) are key localities for reconstructing the ice sheet history during the middle part of MIS 3 because the cave sediments have enabled exceptional accurate dating of the ice margin fluctuations. The caves are found in vertical coastal cliffs in crystalline rocks; they are almost horizontal, up to 100 m long, and formed by wave action during an earlier ice-free period. Sediment records from these caves can be utilised to make inferences of the ice sheet fluctuations because when the ice margin expanded westwards and blocked their entrances, then ice-dammed lakes formed in the caves and laminated clay and silt accumulated on the floor. In contrast, when the caves were open, frost-wedged blocks fell from the roof and, at least during the Ålesund interstadial, ten-thousands of bones were brought into the caves by polar foxes. An accurate chronology is established in the following ways. The Laschamp palaeomagnetic excursion was found in the glaciolacustrine sediments below the Ålesund interstadial strata. This excursion could be correlated with peaks of ^{10}Be and ^{36}Cl in the Greenland ice cores (Mangerud et al., 2003), demonstrating that an advancing ice sheet margin passed by the caves during the Greenland interstadial (GI) 10, or about 41 ka in the GICC05 timescale (Svensson et al., 2008). As many as 32 AMS ^{14}C dates were obtained from very well preserved bones from the Ålesund interstadial beds, yielding ages in the range 34–28 ^{14}C ka (Mangerud et al., 2010). These dating results could also be correlated with Greenland ice core $\delta^{18}O$ curves via marine cores collected close to Greenland (Voelker et al., 1998, 2000), indicating that Ålesund interstadial corresponds with the period of GI 8–GI 7, that is, 38.2–34.5 ka in the GICC05 chronology when using 1950 as the zero year. However, there were fewer dates from the colder Greenland stadial (GS) 8 between GI 8 and GI 7, suggesting that it was colder also in western Norway during this cold spell, but the ice margin did apparently not expand across the caves during GS 8. The inferred age of the end of the Ålesund interstadial was confirmed by the occurrence of the Mono Lake excursion in the glaciolacustrine sediments above the Ålesund bone layer. In the Greenland ice cores, the Mono Lake excursion is identified within GS 7, shortly after the end of GI 7.

Reindeer antlers were among the animal remains that were found in the Ålesund interstadial layers, and the presence of this animal indicates that at a broad zone along the coast was ice-free and had enough vegetation for grazing. Finds of assumed Ålesund interstadial age exist from several sites in Norway, but in no other place is the chronology as well established as in the mentioned caves. At Jæren further south-west along the coast (Fig. 22.6), strata that was termed Sandnes interstadial was in fact identified before the Ålesund interstadial was defined (Feyling-Hanssen, 1974; Raunholm et al., 2004). Olsen et al. (2001a,b) obtained a large number of ^{14}C dates of Mid-Weichselian age from Norway. However, many of these are performed on bulk-sediment samples with total organic content $<1\%$, some even $<0.1\%$, and the reliability of samples with such low carbon content is questionable. Olsen et al. (2001a,b) also report a number of similar ^{14}C ages (conventional and AMS) from marine molluscs, although very rarely they have obtained several consistent dates from one section. There are all reasons to assume that some of the ages reported by Olsen et al. (2001a,b) are correct, but presently, we find it very difficult to distinguish reliable dates from those that are more dubious. In Fig. 22.5, we have schematically drawn a very restricted ice sheet extent during the Ålesund interstadial; it may well have been larger. The reconstruction is partly substantiated by published observations from other sites in Scandinavia (Olsen et al., 2002; Mangerud et al., 2010; Wohlfarth, 2010) and also by modelling results (Arnold et al., 2002; Lambeck et al., 2010).

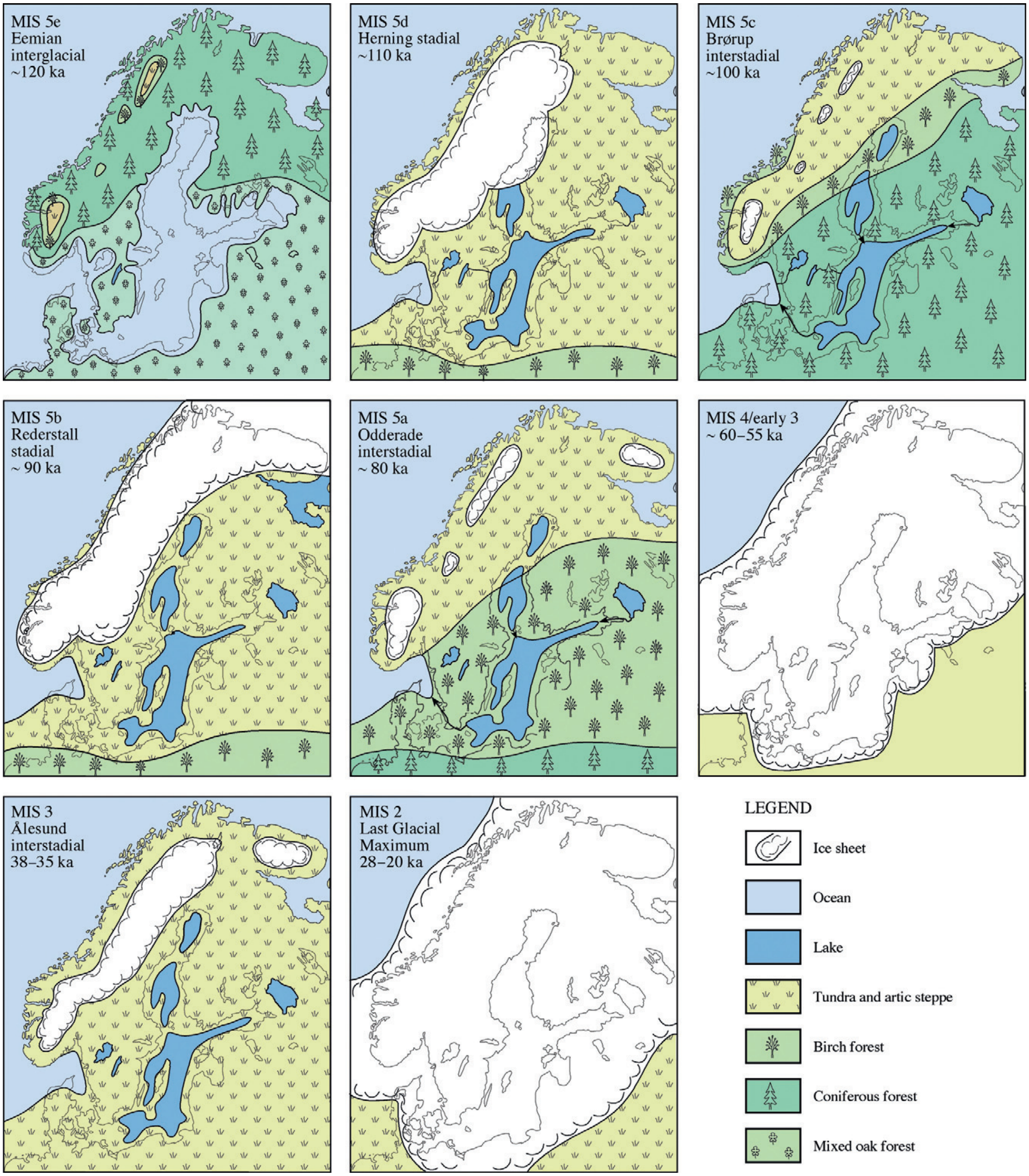

FIGURE 22.5 Conceptual maps illustrating the development of the Scandinavian Ice Sheet and surrounding areas. The original idea and version was by Lundqvist (1992); here, it is developed from Mangerud (2004) and Vorren and Mangerud (2008).

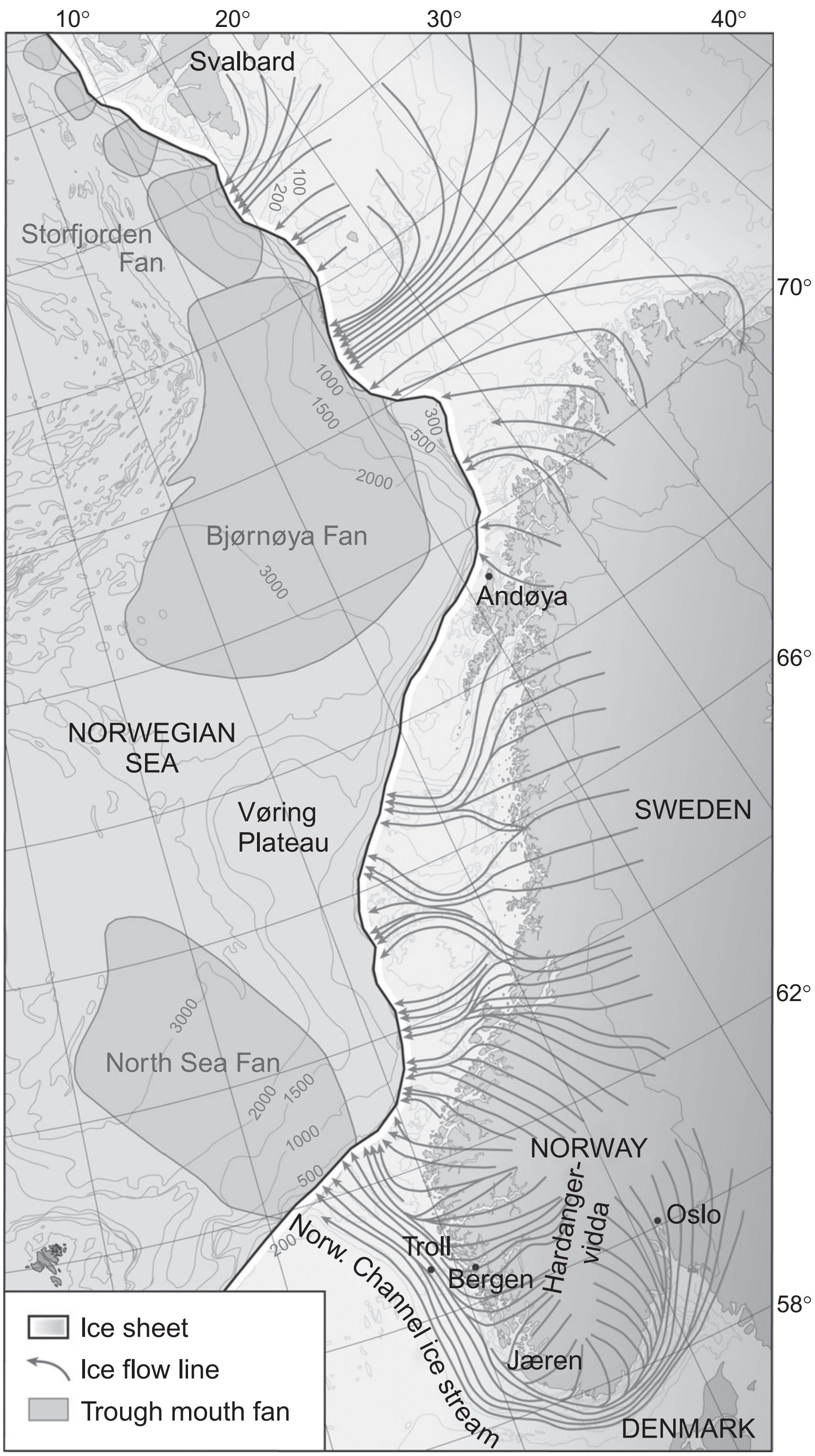

FIGURE 22.6 Reconstruction of ice flow regime of the western margin of the Scandinavian and Barents ice sheets during later phases of the LGM, partly based on Ottesen et al. (2005); here, modified from Vorren and Mangerud (2008). Between the marked ice streams, there was slower-flowing ice. The major trough mouth fans are also marked.

Further back in time, ^{14}C dates become even more uncertain, and their reliability are getting even more questionable. The fact that the caves near Ålesund were blocked in succession from east to west during the Laschamp excursion is therefore important for assessing the finer details of the chronology (Valen et al., 1995; Mangerud et al., 2003) because the implication is that this part of the coast was ice-free shortly before the Laschamp. This older ice-free period is named the Austnes interstadial, and it is supported by some ^{14}C dates from marine shells (Mangerud et al., 2010), but we consider the correlation with Laschamp to be a more reliable dating than the ^{14}C ages. The Austnes interstadial might represent later stages of a long ice-free period starting with the ice margin retreat during the early MIS 3 (Bø interstadial) (Fig. 22.4), or alternatively a shorter ice-free period separated from Bø with one or more ice advance(s).

22.5. THE LATE WEICHSELIAN

22.5.1. The Age of the Last (Local) Glacial Maximum

Here, we use the term LGM in a broad sense, namely as the maximum ice sheet extent in a particular area after the Ålesund interstadial and before about 18 cal. ka. We assume that the maximum extent was not reached simultaneously along the entire coast of Norway. Earlier results were described in some details by Mangerud (2004); here, we will mainly call attention on some more recent findings. Most studies dealing with the Late Weichselian glacial history conclude that there have been (several) fluctuations of the ice margin during the period 30–18 cal. ka BP (Olsen et al., 2001b, 2002; Vorren and Plassen, 2002; Sejrup et al., 2009; Mangerud et al., 2010). Further, marine cores from west of Norway show several peaks of ice-rafted detritus (IRD) that may reflect fluctuations of the ice margin (Dokken and Jansen, 1999; Elliot et al., 2001; Lekens et al., 2009). However, some of the inferred ages for glacial expansions are conflicting; others are difficult to correlate because of uncertain chronology. There seems to be an agreement that the first and largest ice advance from Norway towards south (Denmark) and south-west occurred ~29–27 ka (Houmark-Nielsen and Kjær, 2003; Larsen et al., 2009a; Sejrup et al., 2009), in which case the assumed age (~29 ka) of the Hamnsund interstadial must be wrong (Mangerud et al., 2010). However, Houmark-Nielsen and Kjær (2003) and Larsen et al. (2009a) also find evidence of another re-advance ~23–21 ka from Norway. We emphasise that the maximum extent is not directly dated anywhere north of the mouth of the Norwegian Channel, that is, along most of the ice margin west of Norway, except the Egga II moraine outside Andøya that is dated to about 19 cal. ka BP (Vorren and Plassen, 2002).

22.5.2. The Maximum Extent of the Last (Local) Glacial Maximum

There are no new observations to suggest that the maximum ice sheet extent given in Mangerud (2004) and Svendsen et al. (2004) needs a major revision. However, the glacial morphology of the shelf is now much better mapped, including end moraines and glacial lineations (Ottesen et al., 2001, 2005, 2008; Dowdeswell et al., 2007; Andreassen et al., 2008; Sejrup et al., 2009; Winsborrow et al., 2010); a simplified synthesis map is given in Fig. 22.6. A main result from recent mapping efforts is that a number of fast-flowing ice streams have crossed the continental shelf in between areas of much slower-flowing ice. This would of course also influence the processes and the form of the ice margin. Based on the mentioned mapping, there is now an almost unanimous agreement that the LGM ice sheet reached the shelf edge. This break has for long time been postulated to represent the maximum possible limit because of calving in the deeper water outside.

The only area where there recently has been some debate if the LGM reached the shelf edge or not is around the island Andøya in northern Norway (Figs. 22.1 and 22.11). Vorren and Plassen (2002) mapped two moraines, Egga I and II, along the shelf edge outside Andfjorden just north of Andøya. Their continuation on land was mapped on Andøya and dated to >25 and 19–18 cal. ka BP, respectively (Fig. 22.11). The ages were based on ^{14}C dates from the long and continuous lacustrine records from Andøya (Vorren, 1978; Vorren et al., 1988; Alm, 1993). The reconstructions are reasonable, and they are supported by the exposure dates (although mainly on bedrock) by Nesje et al. (2007). This site demonstrates some of the extreme differences in interpretations, as Lambeck et al. (2002) predicted 1000–1500 m thick ice from isostatic modelling. Considering that the distance to the shelf edge is only 10–20 km outside Andøya, the latter inferred ice thickness sounds unlikely and was also later reduced by Lambeck et al. (2010). We assume that the Egga I (>25 cal. ka BP) advance overran Andøya and reached the shelf edge, partly based on the marine mapping (Ottesen et al., 2005) and partly on the high marine limit of 36 m a.s.l. after Egga I (Vorren et al., 1988); the latter indicating a major glacio-isostatic depression as the eustatic sea level at that time certainly was low.

22.5.3. The Elevation of the LGM Ice Sheet Surface

The existence of possible nunataks during glacial maxima is a more than 100 year's old discussion theme in Norway; a short review is given in Mangerud (2004). Several recent reconstructions of ice thickness are based on glacio-isostatic or glaciological modelling, but here, we will mainly discuss observational data. Geomorphologic features (tors, block fields, trim lines, etc.) have been used to map the upper surface of the LGM ice sheet. In some exceptional cases, that might be right, but it has repeatedly been shown that such forms can survive under (often cold-based parts of) ice sheets (Kleman, 1994; Phillips et al., 2006). We also clarify that any lower limit of such features, including "warm climate clay minerals" (Roaldset et al., 1982) would rather represent the highest level of any Middle- to Late Quaternary glaciation and not necessarily the LGM.

The introduction of exposure dating methods using cosmogenic isotopes (e.g. ^{10}Be and ^{26}Al) opened new possibilities to solve some of the problems. However, for several reasons, it has been a difficult task to obtain reliable dates from isolated mountain summits that potentially were nunataks—or alternatively covered by ice—during LGM: (1) there have often been minimal glacial erosion, as shown by, for example, the comparison of ^{10}Be exposure ages from exposed bedrock surfaces and erratics (Goehring et al., 2008). (2) Erratics may, as the block fields themselves, have survived glaciations after deposition. (3) Erratics from local bedrock or upstream peaks will, because of the limited glacial erosion, more often than erratics from lower terrain have an "inherited age" from earlier exposures. Such peak-source erratics will dominate if the summit is covered by only thin ice. Erratics picked from lower terrain and transported up-hill to a summit are less vulnerable for inheritance, but they would require a thick ice over the summit. Some of the mentioned problems can be solved by using more than one isotope (Fabel et al., 2002), and by obtaining internally and geographically consistent results both concerning exposure dates and glacial reconstructions.

There are few published studies using exposure dating to estimate maximum ice sheet surface elevation near the ice divide. Goehring et al. (2008) present results from two summits, Blåhø and Elgåhogna, located in southern central Norway, slightly north of the assumed ice divide during the LGM (Fig. 22.7). The summit of Blåhø is 1617 m a.s.l. and based on their results, especially a ^{10}Be age of 25.1 ± 1.8 ka from a boulder near the summit; they conclude that the mountain was entirely covered during LGM. Elgåhogna is 1460 m a.s.l., and the results are not unambiguous, but again, based on young ages near the top, they conclude the mountain was completely covered by ice.

Mangerud (2004) considered that the ice sheet probably covered (almost?) all mountains in southern central Norway, and that the ice divide was $>$2000–2500 m a.s.l., before the Norwegian Channel Ice Stream commenced to operate and lowered the ice surface. We maintain that a low ice surface in central southern Norway as postulated by Nesje and Dahl (1992) is difficult to combine with other

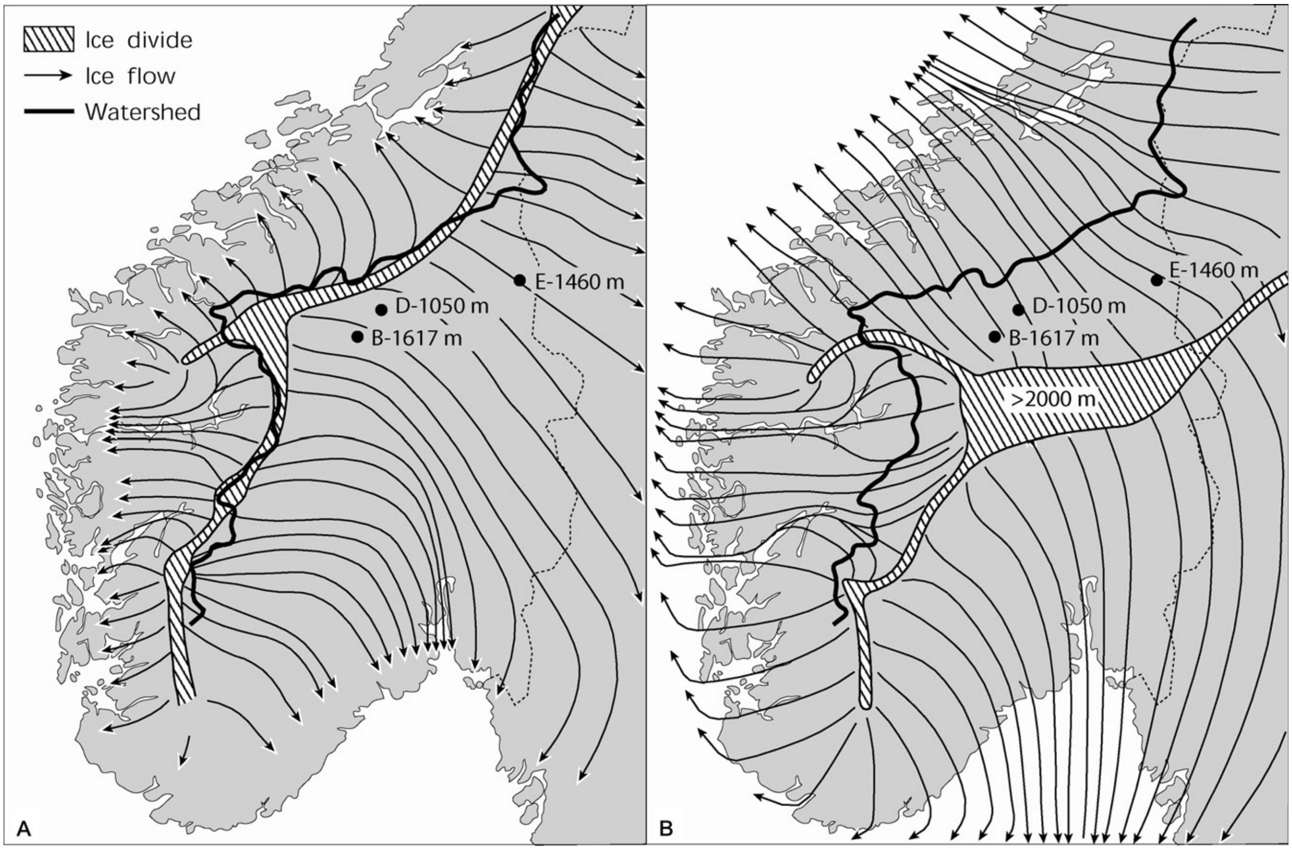

FIGURE 22.7 Conceptual reconstructions of (A) an early phase of a glaciation when the ice divide was located close to the highest mountains and the water divide, and (B) a time around the Last Glacial Maximum when the ice divide was located well south and east of the watershed. Locations and elevations are shown for the summits Blåhø (B in map) and Elgåhogna (E), where exposure dates have been performed and marked, and also for the Dørålen (D) site. Modified from Vorren and Mangerud (2008).

observations such as the geographical extent of the ice sheet to Denmark–Germany and that ice flowed across the water shed from south to north; and even more difficult when Dahl et al. (2010) postulate a maximum altitude for the LGM ice sheet as low as 1050 m a.s.l. in Døràlen. These low ice surfaces are also contradicted by the ^{10}Be dates mentioned earlier (Fig. 22.7). Glaciological and isostatic ice sheet models produce very simplified ice sheets, but they all produce an ice surface up to 2000–3000 m a.s.l. (Näslund et al., 2003; Forsström and Greve, 2004; Peltier, 2004; Siegert and Dowdeswell, 2004; Lambeck et al., 2010), although most allow the highest peaks to project through the surface as nunataks.

Another type of ice-free summits that have been proposed is mountains along the fjords on the west coast. As stated in Mangerud (2004), nunataks in these areas are more feasible from a glaciological point of view than peaks located far inland because the deep fjords would efficiently drain the ice and thus keep the ice surface low. Dahl (1955) provided a strong theoretical argument in favour of such nunataks. He pointed out that the ice sheet margin would calf at the shelf edge because the water depth there increased fast, and that the distance from land to the shelf edge is very short outside Møre (around Ålesund, Fig. 22.1) and Andøya (Fig. 22.1). He further argued that with reasonable ice surface profiles, some of the coast-near mountain peaks would penetrate above the ice surface. This argument is even stronger today when we assume there were fast-flowing ice streams with lower gradients across the shelf. The two mentioned areas are in fact the most studied and discussed areas during the past decades. All scientists who have worked in Nordfjord, just south of Møre, have concluded that, for example, the mountain of Skåla (1848 m a.s.l.) probably remained ice-free, partly based on trimlines and weathering limits and partly on results of exposure dating (Nesje and Dahl, 1992; Brook et al., 1996; Goehring et al., 2008). Similar methods were used along a profile from Andøya and inland (Nesje et al., 2007), although as already mentioned, most exposure dates there were from bedrock exposures. In both areas, the conclusions are geologically reasonable, but not unambiguously demonstrated.

22.5.4. The Deglaciation

A new reconstruction of the deglaciation of the Eurasian ice sheets, based on a large database of radiometric dates, geomorphologic elements and stratigraphy, is presently under construction (Gyllencreutz et al., 2007). Here, we will describe some few examples of deglaciation moraines, chronology and processes.

The deglaciation of Norway can be subdivided into three periods. The first period starts when the ice margin withdraws from its maximum position and lasts until the onset of the Younger Dryas cold reversal (Fig. 22.1). During this first period, the ice sheet melted and calved back so that the continental shelf became completely ice-free and also parts of the coast became deglaciated. Obviously, this ice retreat led to a major thinning of the remaining ice sheet further inland. No end moraines or other ice-marginal features from this period have been mapped over long distances. The next period is the Younger Dryas when the ice margins re-advanced in some areas, notably in the Bergen area (see below), halted in others and even withdrew in some areas as, for example, in the area around Trondheim (Fig. 22.1; Reite, 1994). Moraines from the Younger Dryas are the only deglacial moraines that can be mapped continuously around the Scandinavian Ice Sheet (Fig. 22.1), although the moraines were not formed strictly synchronously (Mangerud, 1980). The third period is the Early Holocene, when the ice sheet disappeared during some 1000–1500 years. There were some distinct halts in most areas, often depending on topographical situations as sills in fjords or palaeo-fjord heads.

22.5.5. The Norwegian Channel and Oslofjord

The Norwegian Channel Ice Stream had a different pattern than the other ice streams on the continental shelf (Fig. 22.6). The ice stream flowed southward in Oslofjorden, continued around the southern coast of Norway and collected ice flowing radially from the SW mountain plateau (Hardangervidda) and ended with a northward flow at the mouth of the channel. The ice stream was not active during the entire Late Weichselian. As discussed in Mangerud (2004), the first MIS 2 ice flow to reach Denmark crossed the Norwegian Channel and transported erratics from Norway to Denmark. This development was recently substantiated by Larsen et al. (2009a) and Houmark-Nielsen (2010). According to Mangerud (2004), the fast ice flow probably started at the mouth of the Norwegian Channel and subsequently propagated up-ice to form a long ice stream. The main point here is that the ice stream was active during the later phases of LGM and that deposition of glacial sediments at the mouth of the channel ceased at 19.0 ± 0.2 cal. ka BP (King et al., 1998; Nygård et al., 2007). Some 220 km upstream the well-dated Troll cores (Fig. 22.6) indicate a deglaciation near 18.5 cal. ka (Sejrup, et al., 2009) implying very fast calving up the deep channel, whether this etreat was triggered by a sea level rise or a climate warming. The oldest obtained ^{14}C ages that show ice-free conditions on land in southern Norway were obtained from Jæren (Fig. 22.6) where several dates indicate a deglaciation between 17 and 16 cal. ka (Knudsen, 2006). At that time, the Norwegian Channel must have been ice-free this far upstream.

Around the Oslofjord, the pattern of ice retreat is known in some more detail. As seen from Fig. 22.8, a major calving

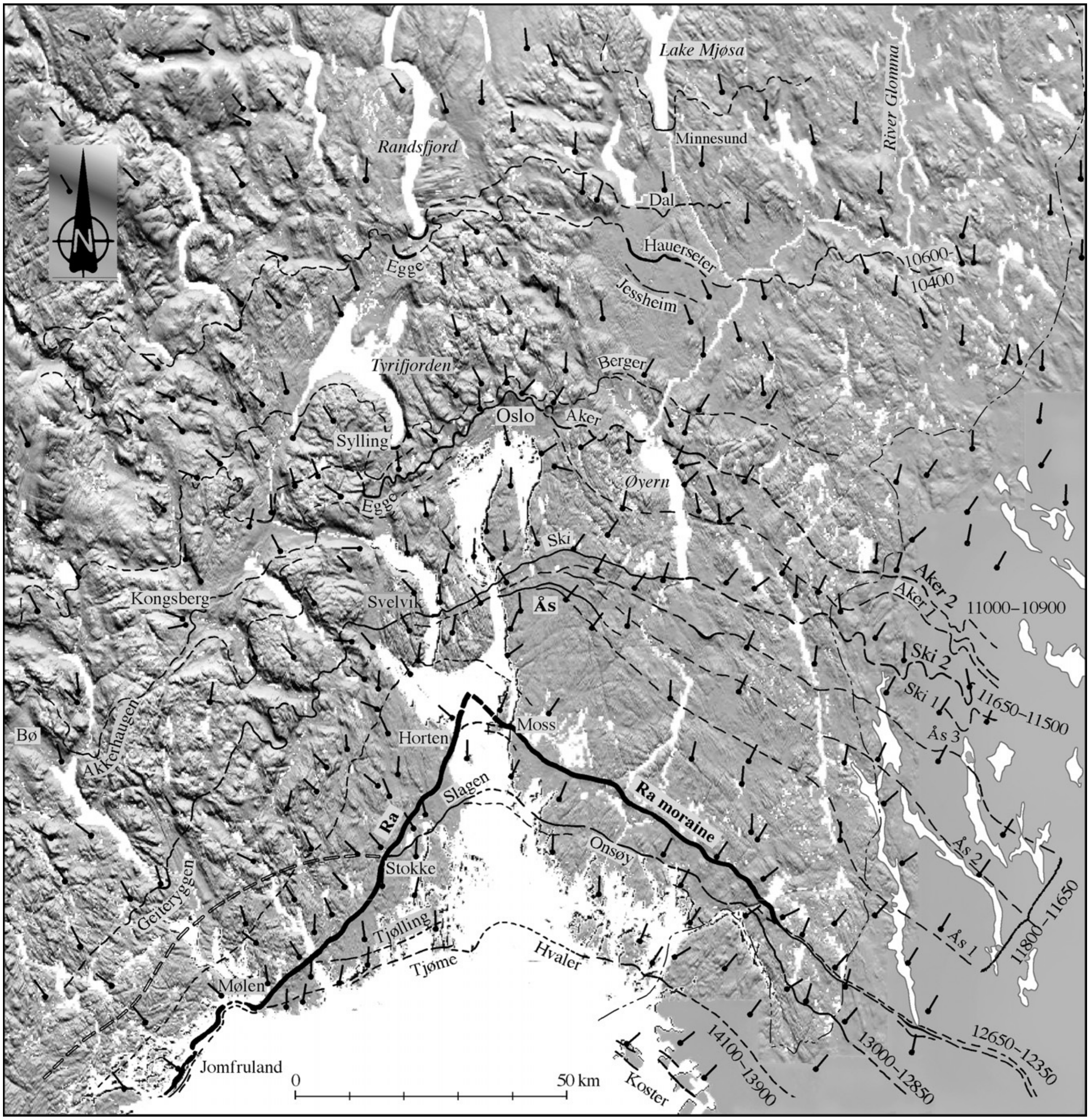

FIGURE 22.8 A map showing ice-marginal deposits and directions of glacial striae around the Oslofjorden. Compiled by Rolf Sørensen (unpublished); here, modified from Vorren and Mangerud (2008). Ages are given in calibrated years BP, mainly based on calibrated ^{14}C dates. Stippled lines indicate less distinct moraines or uncertain correlations.

bay developed along the axis of the deeper part of the fjord, where the main ice stream was flowing some few thousand years earlier. The oldest moraine that can be correlated across the fjord is the Tjøme–Hvaler moraine formed about 14 cal. ka BP (Sørensen, 1979) and correlated with Trollhättan moraine in Sweden (Lundqvist and Wohlfarth, 2001). It should be noted that on the eastern side of the fjord, the younger moraines are oriented in a more or less parallel retreat fashion, whereas on the west side the prominent (Younger Dryas), Ra moraine cut across the older moraines, reflecting a re-advance in this area (Bergstrøm, 1999). Mainly, based on ^{14}C dates of marine molluscs, the prevailing view has been that the Ra moraines around Oslofjorden were deposited during the middle part of the

Younger Dryas, whereas the Ås–Ski moraines were formed during later parts of the Younger Dryas (Mangerud 2004). However, some new dates obtained by Rolf Sørensen (written communication December 2009, Sørensen et al., in press) may indicate that only the Ra moraines are of Younger Dryas age, and that the Ås–Ski moraines are younger. If correct, this would make the age of the Younger Dryas re-advance on the west side of the Oslofjord more in line with the re-advance in the Bergen area (see below) that originated from the same ice culmination over the mountain plateau Hardangervidda (Fig. 22.6).

The Ra moraines consist for a large part of diamicton, mainly till, except for some meltwater deposits that are localised near the valley mouths. The glaciofluvial components increase in the younger moraines, but ridges of till are found also in the Ås, Ski and Aker moraines. The even younger and large ice margin deposits at Berger, Jessheim, Hauerseter, Dal and Minnesund (Fig. 22.8) are totally dominated by glacifluvial sediments. The marine limit around Oslo is 220 m a.s.l., and in valleys distal to the mentioned ice-marginal deposits, there are up to 100 m thick deposits of glaciomarine clay.

22.5.6. Western Norway

The break-up of the Norwegian Channel Ice Stream must have had a major impact on the ice sheet configuration and ice flow over western Norway. If we consider the Bergen area, the westerly ice flow over land turned northwards into the Norwegian Channel (Fig. 22.6) before the ice masses ended up as icebergs at the shelf edge some 220 km further to the northwest. The mentioned distance to the ice margin suddenly became much shorter when the ice disappeared from the northern segment of the Norwegian Channel. The inferred fast retreat from the mouth of the Norwegian Channel to the location of the Troll cores probably resulted in a much steeper gradient of the ice surface towards the mountain areas further inland, that is, to the east in Fig. 22.9. This may have led to a situation where the ice margin during deglaciation was "hanging" on the outer islands west of Bergen (Fig. 22.9), calving into the Norwegian Channel. As mentioned earlier, the Troll site became ice-free as early as 18.5 cal. ka BP, whereas the oldest dates on land (Blomvåg) have yielded significantly younger ages of about 14.6 cal. ka BP (Fig. 22.10). This implies that the ice margin has been calving just outside the coast near Bergen for an extended period of 4000 years (Fig. 22.10), which is the most long-lasting identified halt during the deglaciation of Norway. The standstill was partly a result of ice sheet dynamics, that is, caused by the fast break-up of the Norwegian Channel. However, important for maintaining this position over so long time was also the cool climate combined with the mountain plateaux extending from the central mountains (Hardangervidda) and almost to the coast (Fig. 22.9), together providing large accumulation areas for snow because the equilibrium line was low (Mangerud, 1980). This can be seen when we compare with Jæren (Fig. 22.6), where the high plateaux are located further away and the coast remained ice-free from about 16 cal. ka BP.

During the Bølling, the outermost coast was ice-free for a short period (Fig. 22.10) before the ice margin re-advanced across the outermost islands and presumably again calved into the Norwegian Channel. These outer islands became finally ice-free during early Allerød when the ice margin retreated well inland. However, this retreat apparently stopped before the end of Allerød and was followed by a lasting period with ice growth (Lohne et al., 2007), resulting in an ice advance that reached its maximal position at the very end of the Younger Dryas (Bondevik and Mangerud, 2002). The corresponding glacio-isostatic rebound commenced shortly after (Lohne et al., 2004). During this advance, the ice front reached almost the extreme west coast once again (Fig. 22.9), and the outlet glaciers in the fjords grew to a thickness of 2000 m in areas that were completely ice-free a few hundred years before (e.g. Hardangerfjorden) (Andersen et al., 1995).

At the Younger Dryas/Preboreal boundary, a very fast melting and calving commenced in the fjords. The pattern can be seen from glacial striae which show that the ice flow changed from being regionally directed towards west to flow directions depending on local topography and directed from land areas towards the fjords. The pattern of the moraines (Fig. 22.9) indicates that calving was the dominating process as the distances between the Younger Dryas and Preboreal moraines (Osa–Eidfjord and correlated moraines, Fig. 22.9) are largest along the deeper fjords (Hardangerfjorden and Sognefjorden). Evidently, the retreat rate slowed down considerably at the fjord heads. The Osa–Eidfjord (Anundsen and Simonsen, 1967) and correlated moraines are dated to about 11 cal. ka BP (Bergstrøm, 1975; Mangerud et al., 2009), indicating that the deglaciation of the up to 1300 m deep fjords only lasted some few hundred years during the earliest Holocene. The break-up of the fjords must have led to a major draw down of the entire ice sheet draining towards the fjords.

22.5.7. Andøya–Andfjord

The oldest organic sediments discovered on land in Norway and that are postdating the LGM are found in lake basins on the island Andøya in northern Norway (Figs. 22.1 and 22.11; Vorren et al., 1988). As discussed earlier, they have also been used to argue that Andøya remained ice-free during the LGM, which possibly is correct. In this area, the early deglaciation is well dated and mapped (Vorren and Plassen, 2002). The Bjerka Moraine in Andfjorden has been overrun by ice, but Vorren and Plassen (2002) argue that it

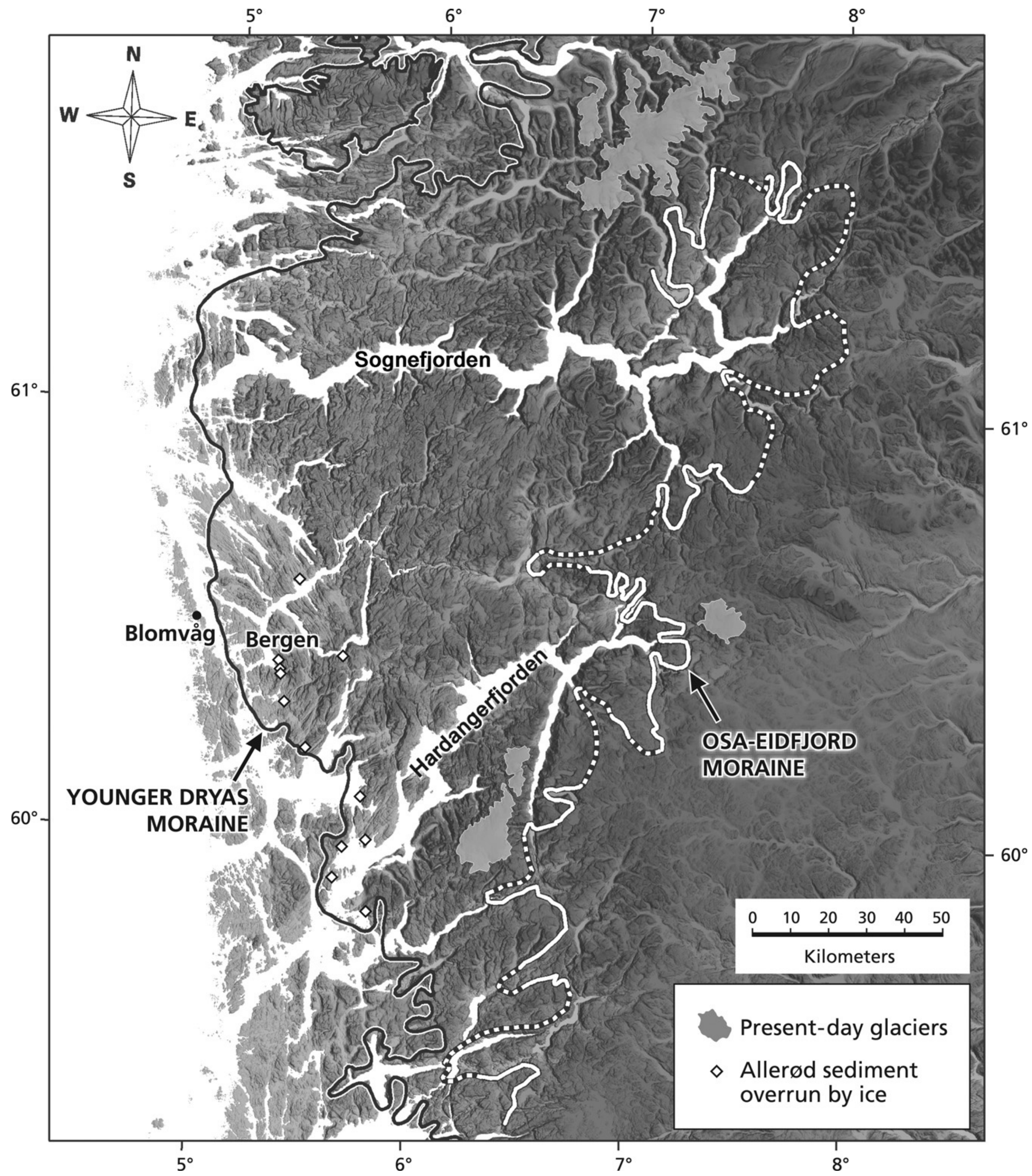

FIGURE 22.9 Map of westernmost Norway, including Bergen (location in Fig. 22.1). The Younger Dryas (Herdla–Halsnøy) moraines are marked. Note an isolated ice cap outside the main ice sheet limits in the upper left corner (Sønstegaard et al., 1999). Some localities around Bergen–Hardangerfjorden with ^{14}C-dated Allerød sediments overrun by the Younger Dryas re-advance are marked with diamonds. The Osa–Eidfjord moraines near the head of Hardangerfjorden and correlated moraines of Preboreal age are also marked. Possible isolated ice caps of Preboreal age outside the Osa–Eidfjord moraines are not indicated.

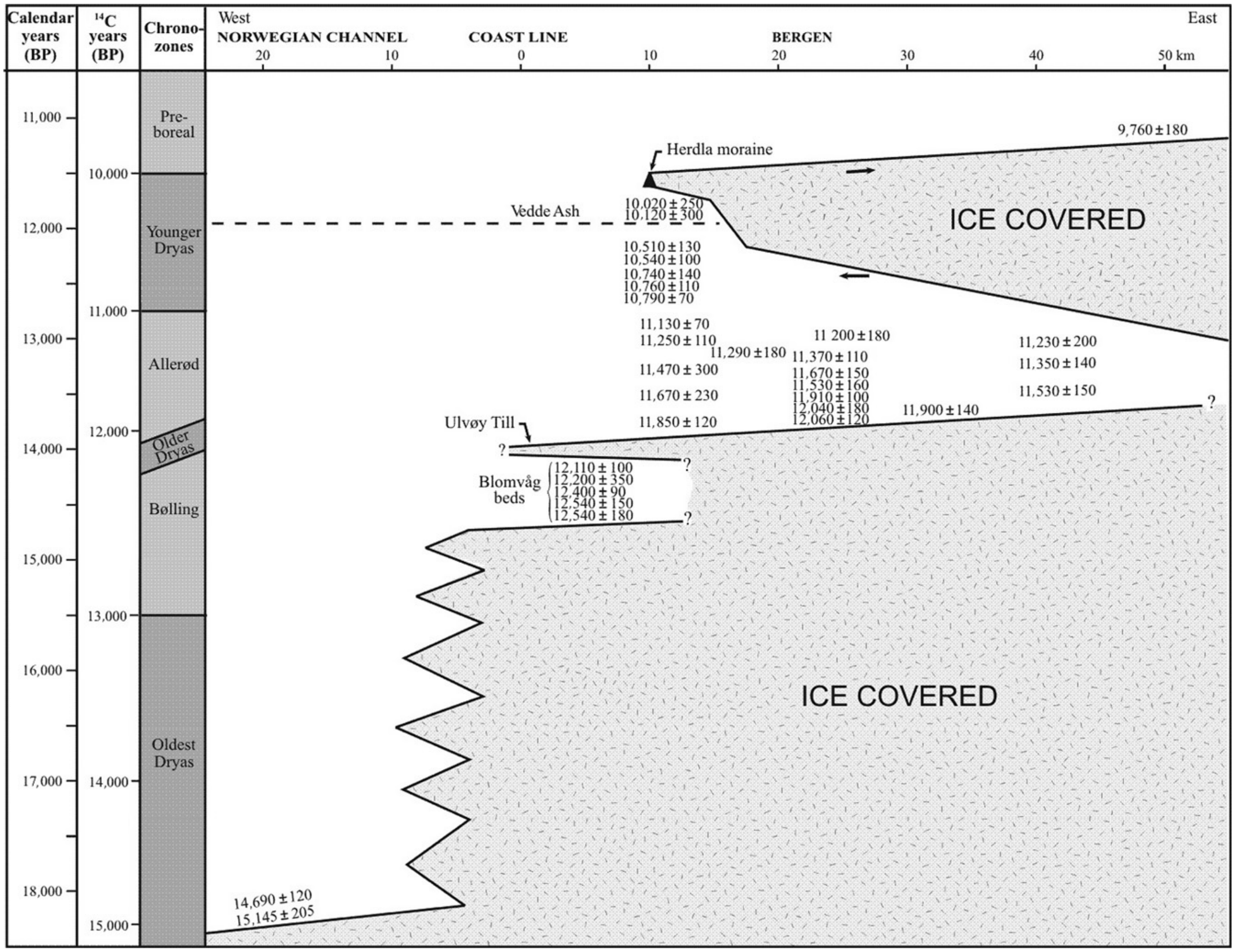

FIGURE 22.10 Time–distance diagram for ice margin fluctuations in the Norwegian Channel–Bergen–Hardanger area. The timescale in calibrated years (Reimer et al., 2009) is linear, but a (non-linear) timescale in ^{14}C years is also shown. Selected representative ^{14}C ages are given in ^{14}C years; for marine samples corrected for a reservoir age of 440 years. The ages plotted in the Norwegian Channel are from the Troll cores (Sejrup et al., 2009), located in the middle part of the channel some 50–60 km from the coast (Fig. 22.1). But in this diagram, these dates are for simplicity plotted in the eastern part of the channel because the deglaciation was from the north with a retreating ice front across the channel. The ages are therefore representative also for the eastern part. Developed from a diagram in Mangerud (2000).

probably has been overrun only once and that its age is between the ages of Egga I and II (Fig. 22.12). If the proposed chronology is correct, then there was a period lasting about 7000 years between two ice margin advances to the shelf edge (Egga I, 26 cal. ka BP and Egga II, 19–18 cal. ka BP), indicating that the ice margin oscillated near the LGM position for an extended period. After 18 cal. ka, there was apparently a fast deglaciation of the shelf and deeper part of Andfjord, until the retreat slowed down when the ice margin reached the coast of about 16 cal. ka. Vorren and Plassen (2002) could not find any end moraine marking the D-event (Figs. 22.11 and 22.12) which is reconstructed along the outer coast; it is marked by a maximum of IRD in cores and a glacial fauna. Moraines of the Skarpnes event, dated to about 14 cal. ka, are found shortly outside or close to the Younger Dryas moraines over large areas in this part of the country (Andersen, 1968), but they are not as continuous as the Younger Dryas moraines. The Tromsø–Lyngen moraines of Younger Dryas age are distinct and well mapped not only in this restricted area but also along most of the coast of Northern Norway (Andersen et al., 1995).

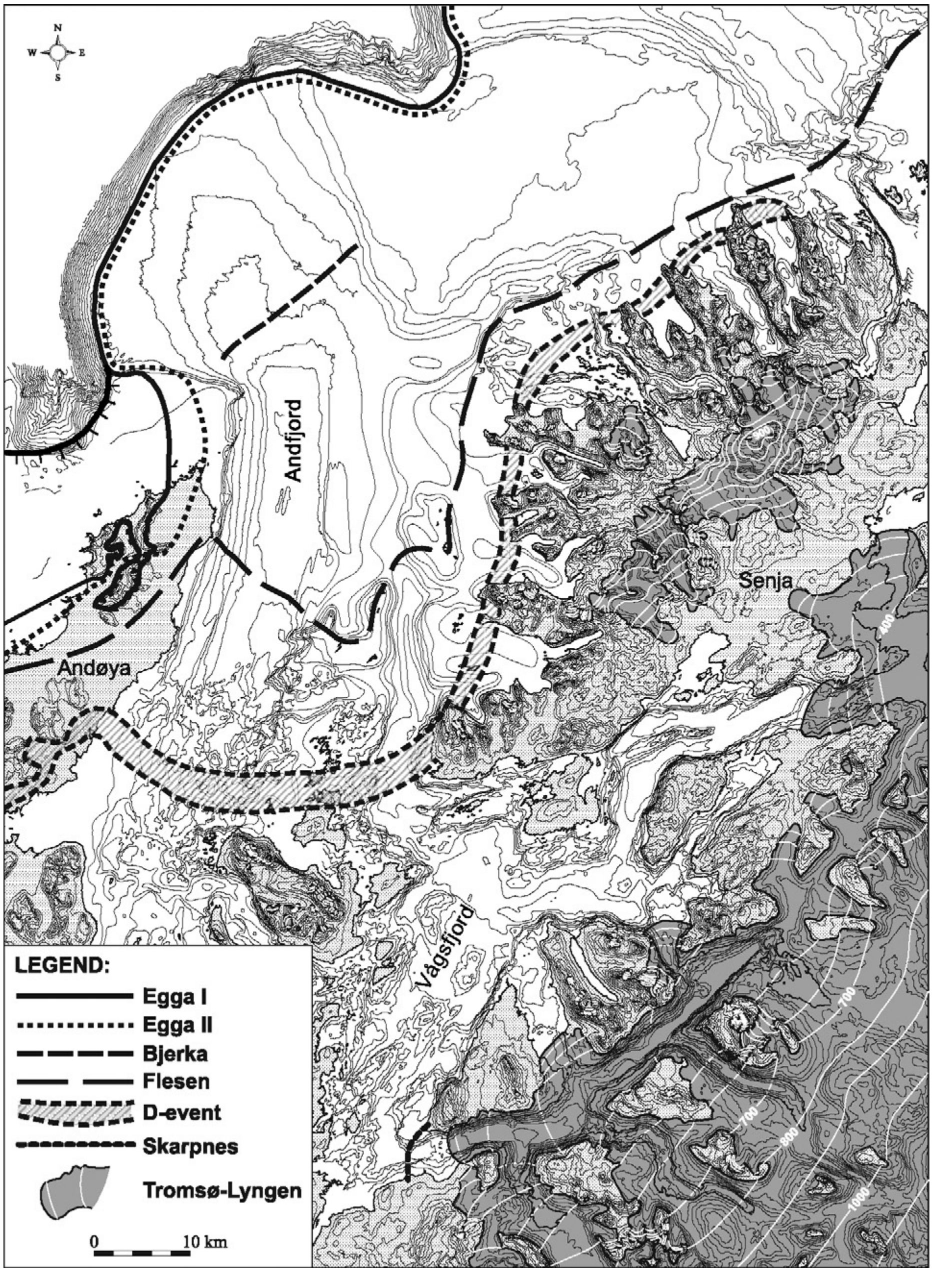

FIGURE 22.11 Glacial map for the area around Andøya–Andfjorden in Northern Norway (location on Fig. 22.1), modified from Vorren and Plassen (2002). Full reconstructions of the ice sheet and glaciers (with white ice surface elevation contours at 100 m interval) are shown for the Younger Dryas. We have added an alternative continuation (marked with tagged line) for the Egga I along the shelf edge towards south (corresponding with the limit shown in Fig. 22.6).

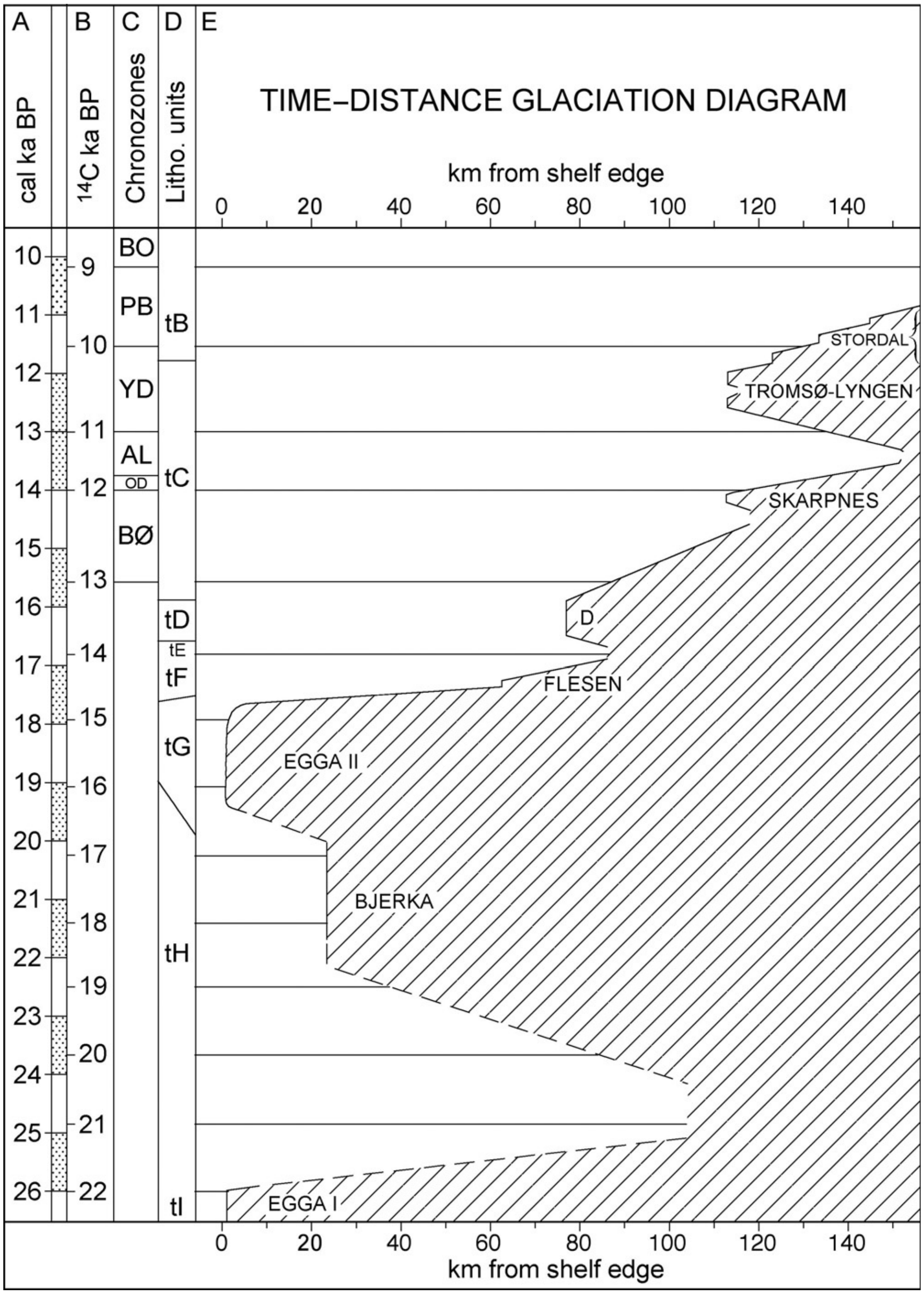

FIGURE 22.12 Time–distance glaciation diagram for the Andfjord area (Fig. 22.11). Modified from Vorren and Plassen (2002). To the left are age scales in both ^{14}C (non-linear) and calibrated (linear) kilo years.

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