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Invited review

Late Cenozoic environmental changes along the Norwegian margin

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ABSTRACT

Our ability to understand the rates and consequences of contemporary climate change is limited by the insufficient duration of instrumental records. Thus, we are not able to fully understand the processes that provide a fundamental control in driving climate changes across different timescales. Palaeo-climate archives, like those preserved offshore Norway, provide our only real window through which to observe long-term rates and styles of climate change. This paper reviews the extensive geological and geophysical data available from the late Cenozoic Atlantic margin of Norway. Along the margin, periods of erosion and deposition have been controlled by agents including fluvial, glacial, and oceanographic processes. Current-controlled sedimentation along the margin provides insight into the connection of the Arctic and Atlantic Oceans from the Miocene onward. Plio-Pleistocene shelf edge progradation of up to 150 km can be linked to the grounding of ice sheets on the continental shelf through observations of buried grounding-zone wedges and mega-scale glacial lineations. The margin is also important for understanding the ability of glaciation to cause topographic relief changes and generate offshore geohazards such as the Storegga Slide, which mobilised some ~3000 km^3 of sediments during the Holocene. Whilst the processes operating along the Norwegian margin are well-understood through the late Cenozoic, there is little geochronological control with which to constrain the environmental changes that have been observed. Concomitant with the wealth of knowledge and the extensive data that are currently available, we propose that the Norwegian margin is an ideal location to be considered for future ocean drilling. The observations of multiple processes, acting independently and together, means ocean drilling could yield information of global significance due to the bridging location of Norway's Atlantic margin between the Arctic and lower latitudes.

1. Introduction

As the Arctic warms at unprecedented rates that are faster than other parts of the globe (Jeffries et al., 2015), human populations are left with the dilemma of how to mitigate against the consequences of environmental change (IPCC, 2014). An obstacle to such endeavours is that instrumental records are often of insufficient length for us to fully understand the processes that provide a fundamental control on any trajectory of change. As such, palaeo-climatic records like those preserved offshore provide important archives through which to observe the rates and styles of long-term climate changes, particularly those beyond the last glaciation.

Extensive industry activity along the Norwegian margin (Fig. 1a, b) has led to the collection of a large database of geological and geophysical records, which provide the opportunity to develop near-continuous spatio-temporal stratigraphic and geomorphological records. These datasets have been used to develop models of how the Norwegian margin has evolved in relation to environmental changes through the late Cenozoic. Here we present an overview of the state of knowledge on the Atlantic margin of Norway, and the cumulative long- and short-term history of environmental changes.

The large number and breadth of factors that have influenced margin development and climate evolution necessitate considering the topics in several sections. In section 2 we describe the geological and oceanographic setting of the Norwegian margin. Section 3 reviews the late Cenozoic evolution of the margin, with a particular emphasis on the development of the glaciogenic stratigraphy and its relation to glacial-interglacial cycles. In sections 4 and 5 we discuss the oceanic and glaciological history offshore Norway and how they are related to the regional evolution of the North Atlantic. In section 6 we review the state of knowledge and highlight future research needs and the case for ocean drilling on the Norwegian margin, before concluding in section 7.

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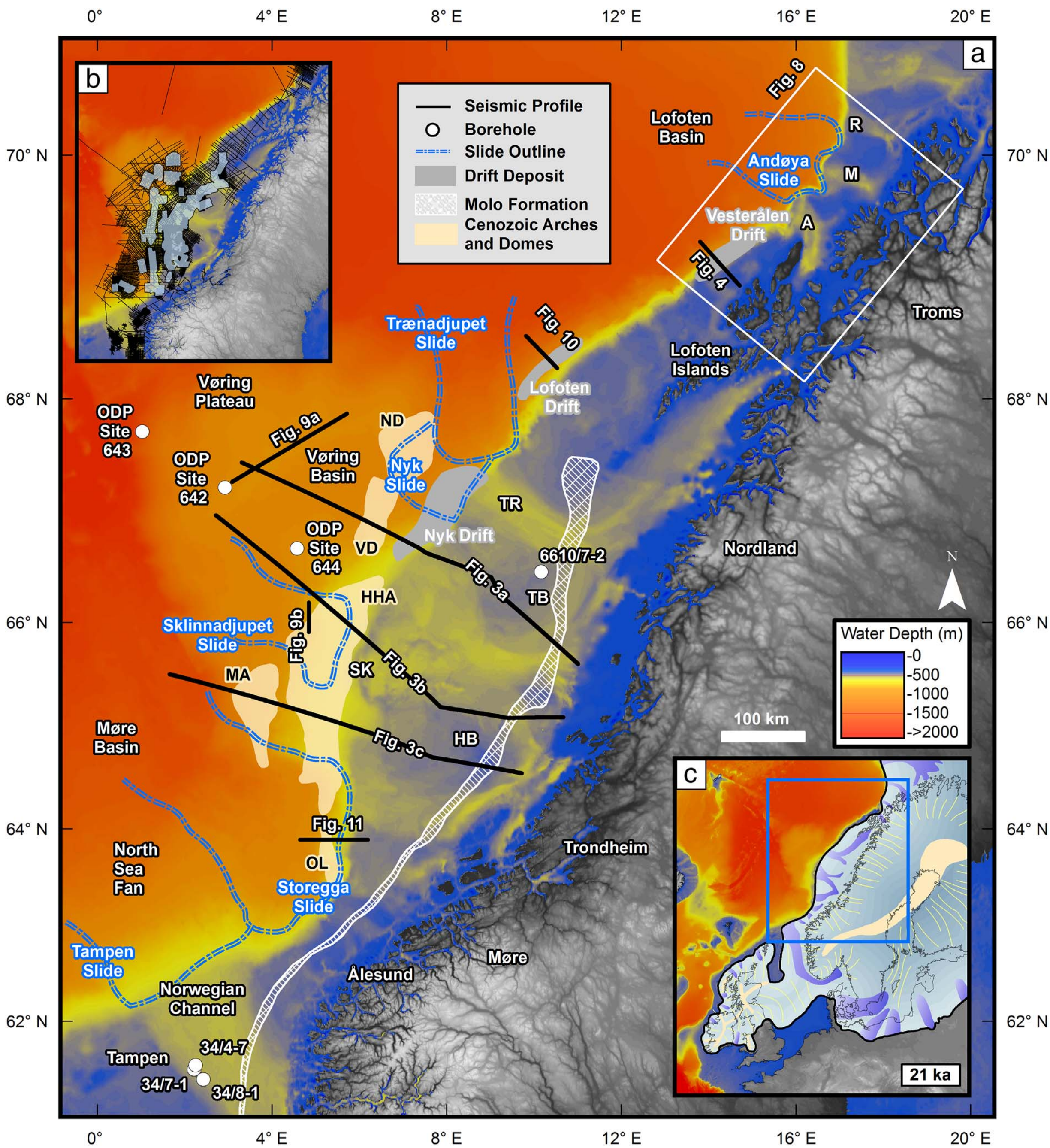

Fig. 1. a) Map showing seafloor topography of the Norwegian margin. The locations of boreholes, submarine slides, domes and arches, drift deposits, seismic lines, and localities mentioned in the text are indicated. Abbreviations are as follows: Andfjorden (A), Haltenbanken (HB), Helland Hansen Arch (HHA), Malangsdjupet (M), Modgunn Arch (MA), Naglfar Dome (ND), Ormen Lange (OL), Rebbenesdjupet (R), Skjoldryggen (SK), Trænabanken (TB), Trænadjupet (TR), and Vema Dome (VD). b) Thin black lines are 2D seismic data, and the blue semi-transparent polygon shows the location of 3D seismic reflection data along the margin. c) Most-credible reconstruction of extent of the last European Ice Sheet at 21 ka from Hughes et al. (2016). Purple-coloured polygons are ice streams compiled from (Punkari, 1997; Stokes and Clark, 2001; Houmark-Nielsen, 2003; Ottesen et al., 2005a; Kleman et al., 2008; Clark et al., 2012). Beige polylines and polygons are inferred flowlines and ice divides from the same sources. The precise timing of each ice stream is uncertain and it is not clear if they all operated at the same time.

2. Study area

2.1. Geological setting

In this review, we primarily concentrate on the Atlantic margin of Norway between 64° N and 71° N and from the contemporary coastline to the continental slope in the west (Fig. 1). This area is bounded by the Troms area in northwest Norway and the Storegga Slide and North Sea Fan region in the south (Fig. 1a). The shelf varies in width from ~20 km off of the Lofoten Islands, to up to 200 km wide on the mid-Norwegian margin. The slope on the mid-Norwegian margin dips gently by just over 1° to the Vøring Plateau, whilst to the south and north the slopes dip by 3–5° to the Møre and Lofoten Basins (Dahlgren et al., 2002). Along most of the margin the contemporary morphology is characterised by shallow banks and cross-shelf troughs (Fig. 1a), where the troughs mark the former locations of ice streams during the last glacial (Ottesen et al., 2005a) (Fig. 1c). Shelf water depths vary from 200 to 500 m with the shelf break generally in water depths of 300–400 m. Water depths deepen to 1000–1500 m across the Vøring Plateau and up to 3000 m in the Møre and Lofoten Basins (Vorren et al., 1998).

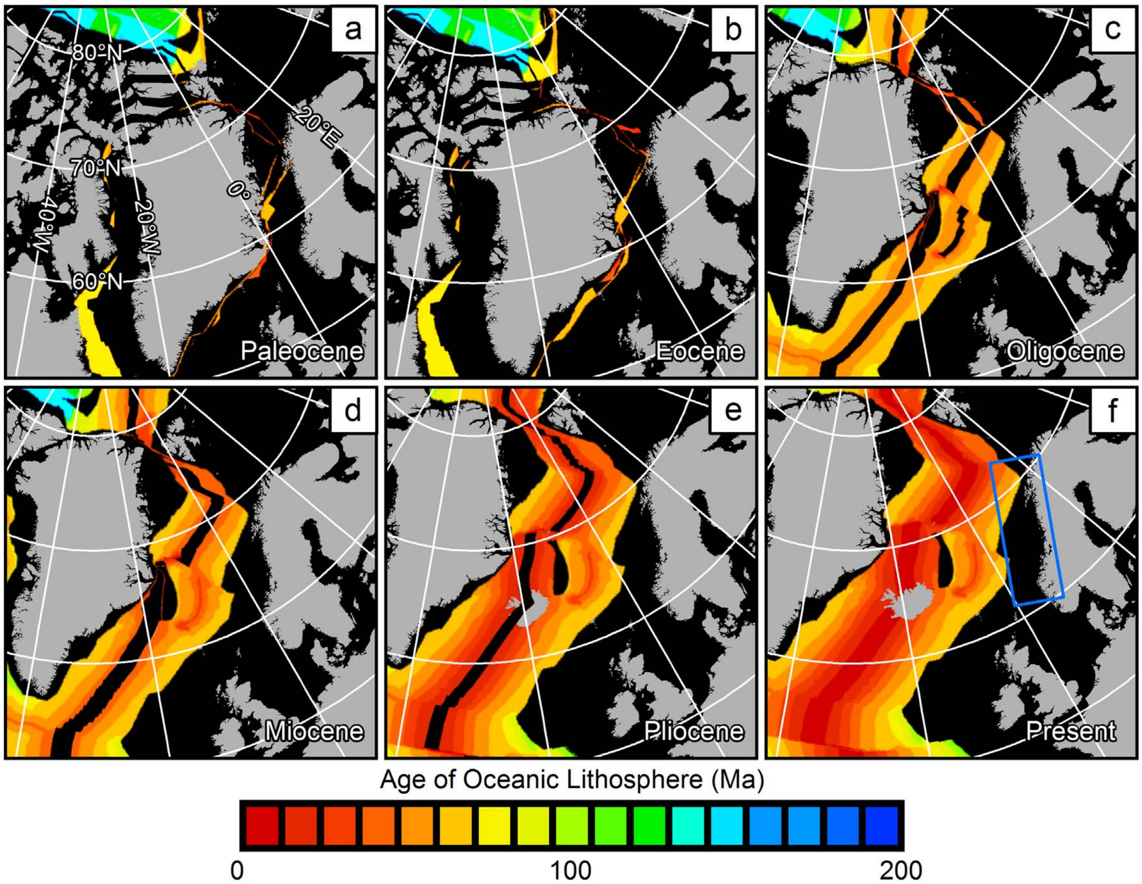

Fig. 2. a–f) Opening of the North Atlantic through different epochs of the Cenozoic era. Colours indicate the age of the oceanic lithosphere whilst the Norwegian-Greenland Sea was opening. Note the narrow seaway in the North Atlantic prior to the Oligocene preventing connection with the Arctic Ocean. As the Fram Strait opened from the Oligocene onwards a deepwater connection was established as circulation patterns more akin to today began to develop. The blue box indicates the contemporary study site investigated in this review. The lines of longitude and latitude that are labelled in panel (a) are the same for all panels. Figure created using GPlates (Müller et al., 2008; Boyden et al., 2011; Seton et al., 2012; Williams et al., 2012).

Mesozoic rifting led to volcanism and the opening of the Norwegian Sea at the Paleocene-Eocene transition (Eldholm et al., 1989; Skogseid, 1994). This continental breakup was associated with uplift of the rifted margins on both sides of the spreading Atlantic, leading to enhanced sediment supply from the elevated areas to the newly formed basins (Doré et al., 1999; Faleide et al., 2010). As the North Atlantic spread through the Cenozoic, the Norwegian margin progressively moved northwards by some 5° of latitude (Fig. 2). As Norway moved north, Cenozoic uplift of mountain chains such as the Himalayas and the opening of the North Atlantic Ocean modified atmospheric and oceanic circulation on a hemispheric scale (Raymo and Ruddiman, 1992; O'Regan et al., 2011). Taken together, these changes led to the general Cenozoic cooling of the climate, which later intensified in the Neogene.

The Brygge Formation was deposited from the early Eocene to the early Miocene and covers large areas underneath the continental shelf (Fig. 3) with the thickest deposits in the Møre and Vøring Basins (Eldholm et al., 1989; Eidvin et al., 2007). It is mainly composed of clay with distal deposits dominated by oozes (Eidvin et al., 2007). Mid Miocene compression caused folding and uplift of several domes and arches in the area (Fig. 1a and 3) that later influenced late Cenozoic depositional patterns (Rise et al., 2005) (e.g. the Helland-Hansen Arch (HHA)). This compressional phase uplifted the landward part of the shelf above sea level and resulted in the mid Miocene unconformity (Løseth and Henriksen, 2005; Smelror et al., 2007; Doré et al., 2008). Above this unconformity the deltaic Molo Formation and its deep water equivalent, the Kai Formation, were deposited from mid Miocene to mid Pliocene across much of the southern and central parts of the Norwegian margin (Henriksen et al., 2005; Eidvin et al., 2007, 2014) (Fig. 3). High resolution seismic data across the Molo Formation shows highly progradational clinoforms dipping up to 10° (Ottesen et al., 2009). Limited core collected from the Molo Formation shows glauconitic sands whilst samples from the Kai Formation show pelagic and hemipelagic clay-rich sediments with siliceous oozes in the basin areas (Eidvin et al., 2013, 2014). After this, the glaciogenic Naust Formation was deposited, with the latest Pliocene and Pleistocene succession measuring up to a kilometre in thickness (Rise et al., 2005; Ottesen et al., 2009).

Whilst this geological setting is typical of the southern and central parts of Norway's Atlantic margin, it is not characteristic of the northern part off of Troms (Fig. 1a). Here, as the Atlantic Ocean opened, the margin became passive, with regional subsidence in the Oligocene (Faleide et al., 2008). Large displacements of basin-bounding faults offshore resulted in a sharply-tapered continental crust and a steeply-dipping continental slope (Fig. 4) (Indrevær et al., 2013). In contrast to the rest of the margin, only a thin Cenozoic sediment cover of siliciclastic deposits is observed (Fig. 4) (Hansen et al., 2012; Tasrianto and Escalona, 2015) pre-dating glaciogenic deposition in the latest Cenozoic (Henriksen et al., 2011; Rydningen et al., 2016).

2.2. Contemporary oceanographic setting

The strength and location of the North Atlantic Current (NAC) is important for the climate of Northwest Europe because it provides an important control on the location of the Arctic Front (Friedrich et al., 2013). The NAC transports heat to high-latitudes in the North Atlantic and likely had an important influence on the onset of the Plio-

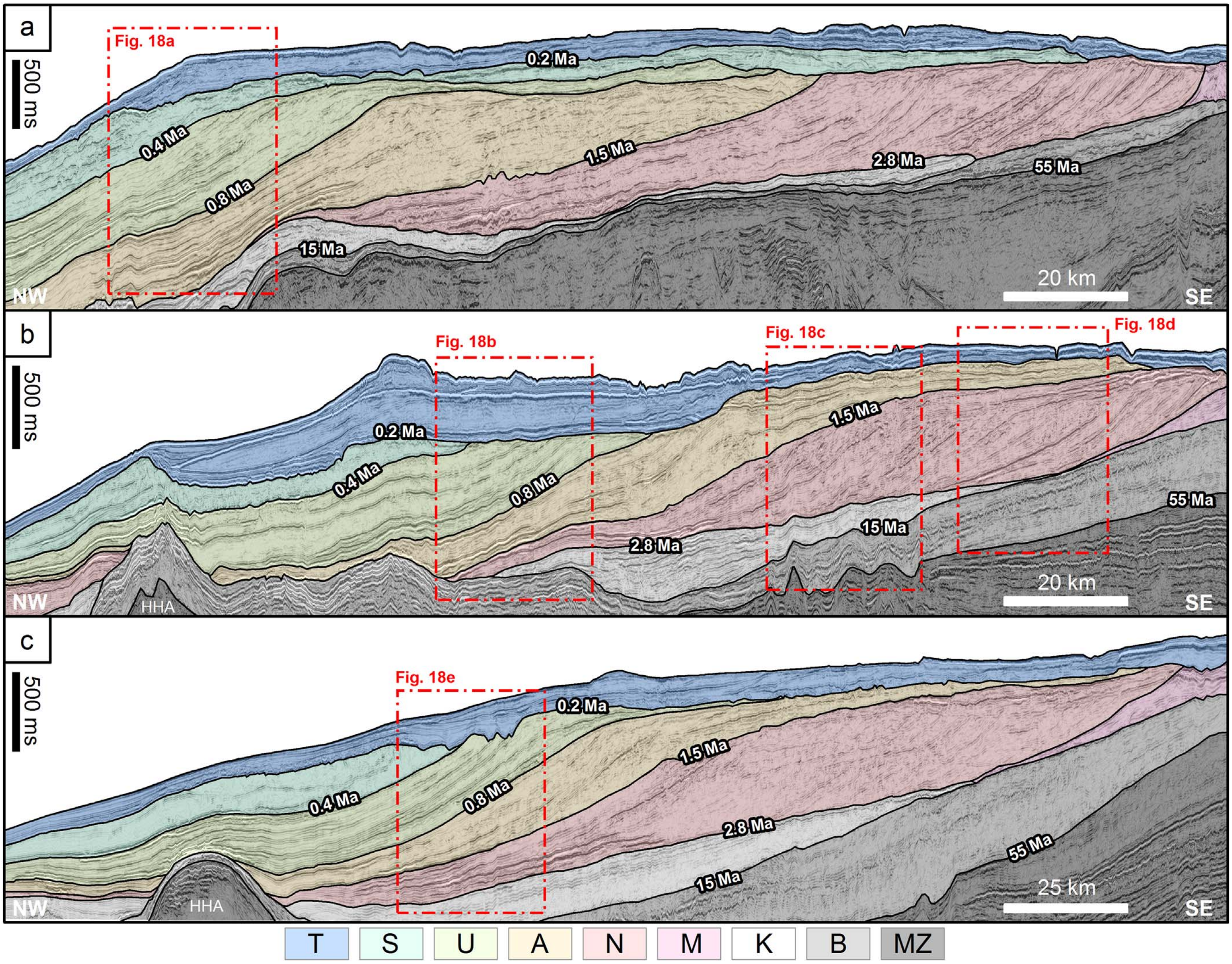

Fig. 3. The Naust Formation (location of the lines on Fig.1a) is composed of late Pliocene and Pleistocene sediments that form a thick succession of prograding sediment wedges and sheet-like units. These are mainly of glacial origin and have been divided into five dated sequences (N, A, U, S, and T) based on the limited availability of high-quality core (Ottesen et al., 2012). Further abbreviations in the stratigraphic key are as follows: Brygge (B), Kai (K), Mesozoic strata (MZ), and Molo (M). Locations of panels in Fig. 18 are indicated.

Pleistocene glaciations (Friedrich et al., 2013). To set the scene for the later discussion of the long-term, late Cenozoic oceanographic evolution along the Norwegian margin, it is important to set it into the context of the contemporary system. The NAC (Fig. 5) is a warm ocean current that is part of the Atlantic Meridional Overturning Circulation (AMOC) (Rahmstorf, 2002; Srokosz and Bryden, 2015). North Atlantic Deep Water (NADW) is produced in the Norwegian and Greenland Seas (NGS) before returning at depth over the Greenland-Scotland Ridge

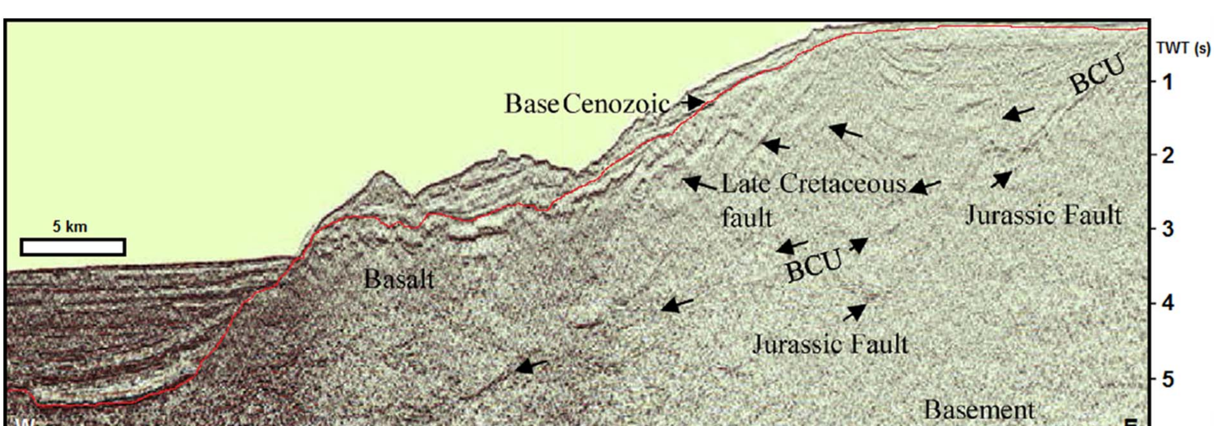

Fig. 4. Seismic profile showing the geological setting of Vesterålen-Andøya on the northern part of the Norwegian Atlantic margin. Red line is the base of the Cenozoic succession. Note that the geological setting has led to a significantly reduced Cenozoic section on the shelf compared to the other parts of the margin to the south. Location of line shown on Fig. 1a. Figure modified from Tasrianto and Escalona (2015).

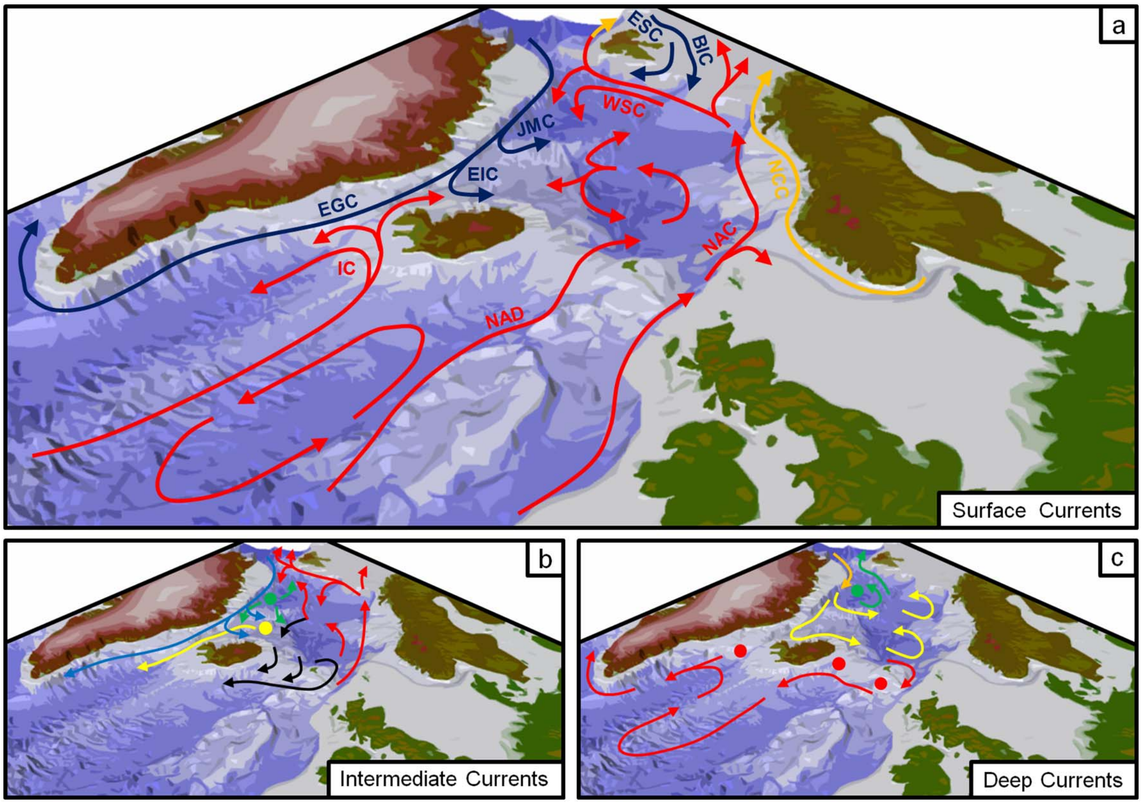

Fig. 5. a) Schematic showing the simplified pattern of surface ocean currents in the North Atlantic. Warm currents are indicated by red arrows and cooler currents by blue arrows. Freshwater, low-salinity currents are shown with the orange arrows. Abbreviations are as follows; Bear Island Current (BIC), East Greenland Current (EGC), East Iceland Current (EIC), East Spitsbergen Current (ESC), Irminger Current (IC), Jan Mayen Current (JMC), Norwegian Atlantic Current (NAC), North Atlantic Drift (NAD), Norwegian Coastal Current (NCC), West Spitsbergen Current (WSC). b) Subsurface currents in the North Atlantic at an intermediate depth. Key flows include those in the Barents Sea and Fram Strait which help to bring water into the Arctic Ocean. Southward flows are also indicated coming through the Fram Strait and across the Greenland-Scotland Ridge. Coloured circles indicate areas of potential slope convection. c) Deep water advection through the Fram Strait and recirculation around the North Atlantic. Coloured circles indicate areas of potential slope convection.
Data compiled from Rudels (2001) and references therein.

(Clark et al., 2002). The remaining NAC surface waters continue into the Arctic Ocean and the Barents Sea through the West Spitsbergen and North Cape currents, respectively (Cokelet et al., 2008; Skagseth, 2008) (Fig. 5). Upon entering the Arctic, Atlantic-derived waters lose upwards of 30% of their heat through ice melt and atmospheric warming (Rudels et al., 2005). The currents then return through the Fram Strait as the colder, less saline East Greenland Current (Telesiński et al., 2014) (Fig. 5).

In the Norwegian Sea, the warm surface waters of the NAC are split into the Norwegian Atlantic Slope Current (NwASC) and the Norwegian Atlantic Front Current (NwAFC) (Skagseth and Orvik, 2002; Skagseth et al., 2004), although the eastern current is commonly referred to as the Norwegian Current or the Norwegian Atlantic Current. The NwASC is near barotropic with velocities of 20–40 cm s^{-1} and flows through the Faroe-Shetland Channel and along the ~500 m isobath in the Norwegian Sea (Hansen et al., 2011) (Fig. 5). It then splits into two branches off of Northern Norway with one current flowing into the Barents Sea and the other north toward Svalbard. Bottom currents associated with the NwASC have been estimated from reworked iceberg scours and show velocities from < 5 cm s^{-1} to over 50 cm s^{-1} (Bellec et al., 2008, 2009).

The NwAFC is a baroclinic current that flows over the Iceland-Faroe Ridge at velocities of 10–15 cm s^{-1} (Hansen et al., 2011). Although its strength and location is less clear compared with the NwASC, the Arctic Front and location of cold Arctic air masses exert an important influence (Hansen and Østerhus, 2000; De Schepper et al., 2015). Northward flux in the eastern NGS can reach up to ~3–7 Sv and is important for water exchange between the North Atlantic and the Arctic Ocean (Aaboe et al., 2009).

The Norwegian Coastal Current flows along the Norwegian coastline before entering the Barents Sea north of Troms (Ingvaldsen et al., 2004) (Fig. 5). The current is buoyancy-driven and consists of low salinity and low temperature water sourced from river runoff into the Baltic and North Seas (Skarðhamar and Svendsen, 2005). Doppler velocities obtained from synthetic aperture radar show that the Norwegian Coastal Current has velocities approaching 40 cm s^{-1} (Hansen et al., 2011).

3. Late Cenozoic margin evolution

3.1. Seismic stratigraphy

The seismic stratigraphy is interpreted using reflection seismic images with Two-Way Traveltime (TWT) as the vertical scale and specific interval velocities are mostly not provided in the literature. Hence we use a standard interval velocity of 2 km s^{-1} to convert time to depth when referring to depositional thicknesses in this section (Ottesen et al., 2009). The uncertainty introduced in this way is likely < 10% and thus equal to or less than uncertainties resulting from poor age constraints and interpolation of thicknesses across regional seismic grids.

After the deposition of the early Eocene to early Miocene Brygge Formation, the Molo Formation and its deep water equivalent, the Kai Formation, were deposited from mid Miocene to mid Pliocene time (Eidvin et al., 2007, 2014) (Fig. 3). The Naust Formation is of latest Pliocene to Holocene age and forms the remainder of the sedimentary succession offshore Norway. It measures up to 1 km thick and consists of a series of prograding clinoforms that extend the shelf edge by up to 150 km to the west and is bounded to the east by the Molo Formation

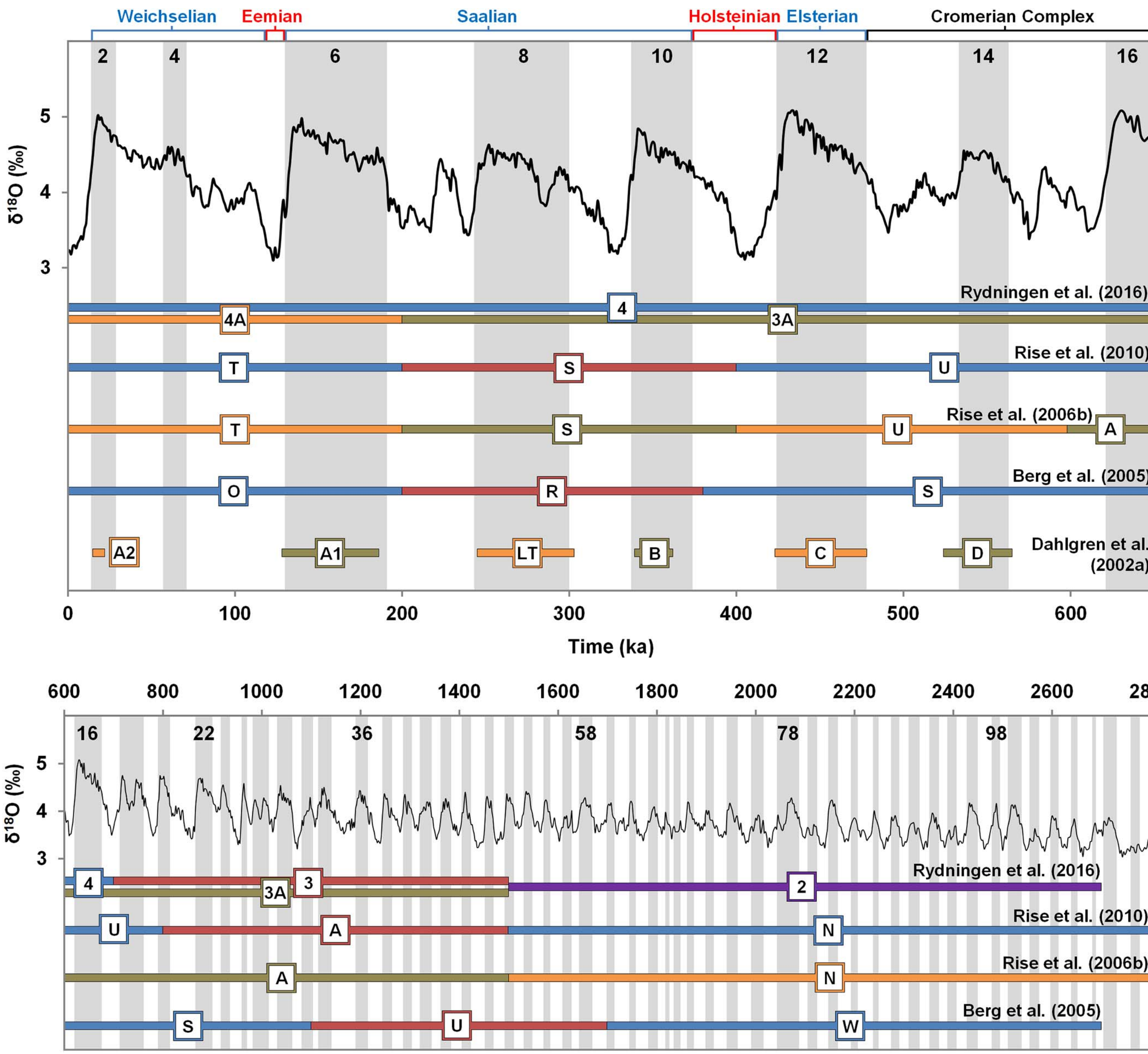

Fig. 6. Comparison schematic of the different chronostratigraphic schemes used on the Norwegian margin that are discussed in the text. Glacial marine isotope stages are indicated by the grey boxes (numbers refer to stage numbers) and the global $\delta^{18}O$ by the black graph. Both are compiled from Lisiecki and Raymo (2005). The glacial and interglacial stage names during the Middle and Late Pleistocene are from the International Comission on Stratigraphy (Cohen and Gibbard, 2011). Note that the Rydningen et al. (2016) scheme has an alternative correlation (e.g. 3A), which is also indicated.

(Eidvin et al., 1998; Dahlgren et al., 2002; Henriksen et al., 2005). The oldest bottomsets of the Naust Formation downlap onto the underlying Kai Formation and thin westwards, where several clinoforms onlap the eastern flank of the HHA (Dalland et al., 1988; Rokoengen et al., 1995) (Fig. 3). Some sediments from the oldest units of the Naust Formation have also been deposited west of the HHA, where they onlap its western flank (Rise et al., 2010). Once the ponded basin to the east of the HHA was infilled, the youngest Naust Formation clinoforms continued beyond the contemporary shelf break and across the HHA. Seismic data here show thin, acoustically-stratified deposits in the distal areas (Rise et al., 2010) (Fig. 3).

The stratigraphy of the Naust Formation and its equivalents has undergone several phases of interpretation, resulting in different terminologies and dates (McNeill et al., 1998; Dahlgren et al., 2002; Berg et al., 2005; Rise et al., 2006b; Ottesen et al., 2014; Rydningen et al., 2016) (Fig. 6). In this review we use the stratigraphic nomenclature of Rise et al. (2010) who divided the main Naust Formation into five units; N, A, U, S, and T from oldest to youngest (Figs. 3 and 6). The two youngest units (S and T) are relatively well-constrained by seismic correlation with borehole data at ODP Site 644A (Dahlgren et al., 2002). However, only a limited number of cuttings with dateable material have been retrieved from the older stratigraphy, meaning that age constraints are limited for the three older units (N, A, and U) (Ottesen et al., 2009).

The N-unit is characterised by up to ~80 km progradation from 2.8–1.5 Ma (Rise et al., 2010) (Fig. 3). Clinoform architecture shows a repeated pattern of low-angle clinoforms that prograde to the northwest, away from the Norwegian mainland. The N-unit appears massive and acoustically transparent, although this is not uniform along strike. Topset preservation is limited due to erosion surfaces related to uplift of the inner shelf and grounded shelf glaciation (Ottesen et al., 2009) (Fig. 3). These unconformities are not continuous and a number of palaeo-shelf breaks are preserved, even when the topsets are truncated. Based on biostratigraphic data, the onset of the glacial deposition that

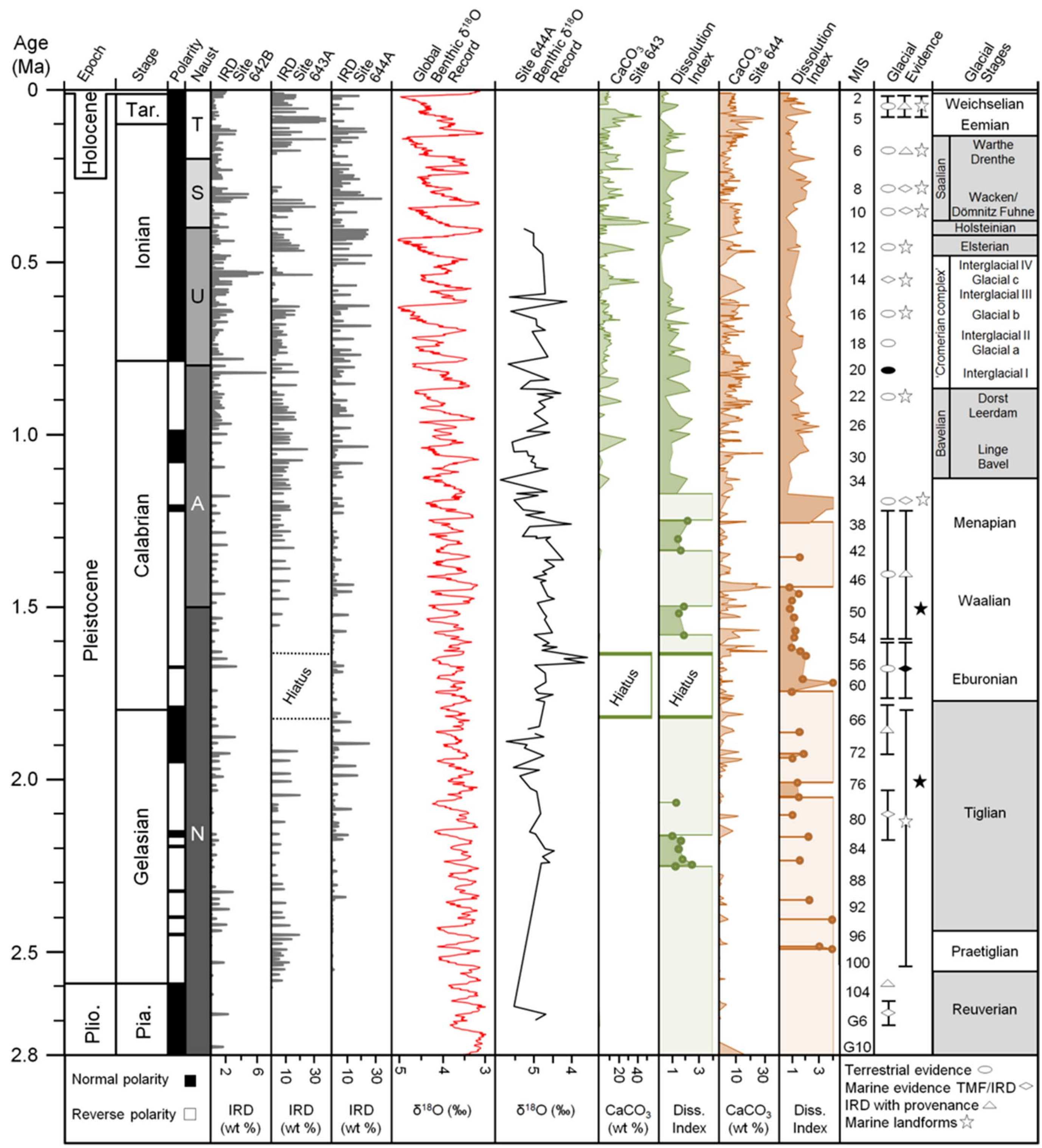

Fig. 7. Summary figure showing the various different geological and geochemical proxies used to infer ice sheet history offshore Northwest Europe. The Naust Formation stratigraphy is taken from Rise et al. (2010). IRD, the local $\delta^{18}O$ record, and $CaCO_3$ data compiled from ODP Sites 642–644 (Jansen et al., 1989; Krissek, 1989; Henrich and Baumann, 1994; Henrich et al., 2002). The global $\delta^{18}O$ is from Lisiecki and Raymo (2005). The glacial evidence section is summarised from a number of studies looking at marine and terrestrial records of ice sheet history in Northwest Europe (Stoker et al., 1994; Baumann et al., 1995; Mangerud et al., 1996; Snoeckx et al., 1999; Huuse and Lykke-Andersen, 2000b, 2000a; Jansen et al., 2000; Scourse et al., 2000; Grousset et al., 2001; Knutz et al., 2001; Dahlgren et al., 2002; Flesche Kleiven et al., 2002; Clark et al., 2004; Hemming, 2004; Hjelstuen et al., 2004b; Rise et al., 2004; Sejrup et al., 2005; Dowdeswell et al., 2006; Ehlers and Gibbard, 2007; Knies et al., 2007; McCabe et al., 2007; Peck et al., 2007; Bradwell et al., 2008; Hubbard et al., 2009; Rose, 2009; Hibbert et al., 2010; Stewart and Lonergan, 2011; Lee et al., 2012; Ottesen et al., 2012; Thierens et al., 2012; Dowdeswell and Ottesen, 2013; Newton et al., 2016). Where dating of this evidence is less reliable, the symbol is coloured black. We have included offshore data from the North Atlantic and evidence from mainly Scandinavia. In addition, evidence from northern Britain has been included because of the similar climatology to that of southern Norway. As a consequence of this, any glacial growth and decay in response to climate changes was likely coeval across this area. The glacial stage names come from Cohen and Gibbard (2011).

comprises the N-unit occurred in the latest Pliocene (Eidvin et al., 2000) and is assumed to be coeval with increased ice-rafted detritus (IRD) measured in ODP boreholes on the Vøring Plateau at ~2.8 Ma (Jansen and Sjøholm, 1991; Fronval and Jansen, 1996) (Fig. 7).

Unit A shows similar acoustic characteristics to the N-unit and covers the time period 1.5–0.8 Ma (Rise et al., 2010) (Fig. 3). Variable progradation along strike allowed the development of a narrower shelf in the region of Møre to the south. Here the clinoforms are steeper and more aggradational compared to the clinoforms in the main sedimentary wedge offshore mid-Norway (Ottesen et al., 2009). The steeply-dipping clinoform architectures and variable sedimentology have contributed to the multi-slide Storegga Complex (Bryn et al., 2005; Solheim et al., 2005) and the succession here comprises a significant amount of slide-derived debrites.

The U-unit was deposited from 800 to 400 ka and is composed of several sub-units separated by high amplitude reflections (Fig. 3). These high amplitude reflections might represent several glacial-interglacial cycles during the Cromerian (Marine Isotope Stages (MIS) 21-13) and may indicate the delivery of coarser-grained material to the shelf edge when ice was grounded on the continental shelf. Dahlgren et al. (2002) observed glacial diamicton (within their Naust Formation D unit, Fig. 6) on mid-Norway, suggesting shelf-based glaciation from at least MIS 14. This finding is in agreement with Berg et al. (2005) who observed glaciomarine deposits (within their Naust Formation S3-S5 units, Fig. 6) that indicate shelf-based glaciation in the Ormen Lange area around this time. The seismic architecture shows a combination of progradation and aggradation with limited rollover preservation due to later glacial erosion and topset truncations. The thickest part of this unit is in the Skjoldryggen area (Fig. 1a) where it is up to 500 m thick (Ottesen et al., 2009).

From 400 to 200 ka unit S was deposited and reaches up to ~350 m in thickness before pinching out in the west, and is characterised by a massive acoustically transparent seismic facies (Rise et al., 2006b; Ottesen et al., 2012) (Fig. 3). A large debris-flow lobe and a number of glaciogenic debris-flow lenses have also been observed in this unit (King, 1993; Dahlgren et al., 2002; Nygård et al., 2003). The unit is deposited beyond the modern shelf edge as two main depocentres after the Sklinnadjupet Slide (Fig. 1a) removed the central part of the unit at about 250 ka, leaving behind a 90 km wide slide scar (Rise et al., 2006b). The age of this unit is well-constrained through correlation of seismic data with an age-model from ODP Site 644A (Henrich and Baumann, 1994; Dahlgren et al., 2002). This shows that Fennoscandian ice extended across the mid-Norwegian shelf to the shelf edge during MIS 10 and to the inner shelf during MIS 8 (Dahlgren et al., 2002). This corresponds with observations from Storegga Slide scar indicating glaciomarine deposition during MIS 10 and ice-distal deposition during MIS 8 (Berg et al., 2005).

The T-unit covers the last two major glaciations from 200 to 0 ka. Across large areas the base of the unit truncates palaeo-shelves of the earlier Naust Formation units (Fig. 3). The truncation surface is called the upper regional unconformity (Rokoengen et al., 1995). Although this surface is generally thought to relate to the ice sheet that deposited the S-unit, across the shelf the unconformity is complex and time-transgressive (Rise et al., 2005; Ottesen et al., 2012). On the inner shelf this unconformity is generally associated with erosion during the last glaciation. The unit is composed of mainly flat-lying units that are generally composed of glacial diamicton (Dahlgren et al., 2002; Dahlgren and Vorren, 2003; Hjelstuen et al., 2004b; Ottesen et al., 2012). These diamicts were deposited as ice sheets extended onto the continental shelf during MIS 6 and 2 (Dahlgren et al., 2002). These units can be broken into smaller, wedge-like units that terminate on the shelf (Dahlgren et al., 2002) and have been described as ‘till tongues’ which are composed of tills inter-fingered with glaciomarine sediments (King et al., 1987, 1991) or as grounding-zone wedges (Dowdeswell et al., 2008). Along the entire margin, where these features occur close to the shelf edge, material was transported by gravity flows and redeposited on the continental slope (Elverhøi et al., 1997; Vorren and Laberg, 1997; King et al., 1998). The main depocentres are in the Skjoldryggen area (Fig. 1a), where the sediments are up to ~400 m thick (Ottesen et al., 2009), and to the south of the main sedimentary wedge where the accumulations are up to ~400 m thick in the North Sea Fan (Nygård et al., 2005, 2007).

Whilst the dates for the earliest unit boundaries of the Naust Formation are poorly constrained (Rise et al., 2010; Ottesen et al., 2012), the dates for the later units are comparatively robust (Dahlgren et al., 2002). This suggests confidence in the dates for the T- and S-units of Rise et al. (2010), which correspond to the A-and B-units of Dahlgren et al. (2002) (Fig. 6). However, the use of glacial stages complicates comparisons because different studies attribute different chronology to different stages. The T-unit is suggested to cover the Weichselian (MIS 5d-2) and Saalian glaciations (MIS 10-6) (Rise et al., 2010; Ottesen et al., 2012). Following the Dahlgren et al. (2002) seismic correlation with ODP Site 644A, the early T-unit corresponds to MIS 6 (late Saalian). Thus, although the S-unit is commonly attributed to the Elsterian glaciation (Rise et al., 2006b; Ottesen et al., 2012), the unit is actually composed of sediments deposited during the MIS 10 and 8 peaks of the Saalian Complex (Fig. 6) (Dahlgren et al., 2002). The Elsterian glaciation occurred during MIS 12 according to the International Comission on Stratigraphy (Cohen and Gibbard, 2011) and Dahlgren et al. (2002) correlate this to their C-unit, which is the upper part of the U-unit of Rise et al. (2006b) (Fig. 6). This discrepancy is an issue of glacial stratigraphic nomenclature, rather than geochronology. It probably relates to previous uncertainty on the age of the Holsteinian, the interglacial preceding the Saalian. The age of the Holsteinian is now correlated to MIS 11 (Cohen and Gibbard, 2011), meaning that the attribution of the Elsterian glaciation to the S-unit is incorrect and that the unit is comprised of sediments delivered during the early Saalian (MIS 8-10), in agreement with Dahlgren et al. (2002). The use of marine isotope stages is clearly a better basis for correlating glacial cycles rather than local glacial stage names. This removes confusion over glacial nomenclature and makes correlation to other glaciated margins much easier.

On the northern part of the margin, less work has been carried out compared to the central and southern parts. Rydningen et al. (2016) has attempted to correlate the stratigraphy here with records from the Naust Formation and the Barents Sea (Rise et al., 2010; Laberg et al., 2011). This has shown that in the late Cenozoic, prior to the onset of major Northern Hemisphere glaciation, several submarine canyons were infilled and then cut by different periods of turbidity current erosion (Rydningen et al., 2016). From ~2.7–1.5 Ma (i.e. Naust Formation N-unit), glaciomarine and glaciofluvial sedimentation is thought to have dominated before ice sheets later reached the shelf edge from ~1.5–0.7 Ma (Naust Formation A-unit) and deposited tills that were later reworked into debrites and turbidites (Rydningen et al., 2016). After 0.7 Ma (Naust Formation units U, S, and T) ice streams were likely present on the shelf through most glacial periods (Rydningen et al., 2016).

3.2. Sedimentary environments

Despite the large number of industry wells, there is relatively little core material available from within the Naust Formation, particularly from the oldest units which have not been penetrated by DSDP or ODP boreholes. Side-wall core data from well 6610/7-2 on Trænabanken (Fig. 1) shows that the lower part of the Naust Formation is composed of fine-grained diamict that is occasionally interbedded with sandy layers up to a few metres thick. The diamict contains angular and rounded gravels derived from sedimentary and crystalline rocks (Ottesen et al., 2009) and are thought to have been deposited in glaciomarine and grounded ice environments, with some sediments being reworked from the Molo Formation or redeposited at the shelf break by gravity-driven processes (Dahlgren et al., 2002; Rise et al., 2005). Wells

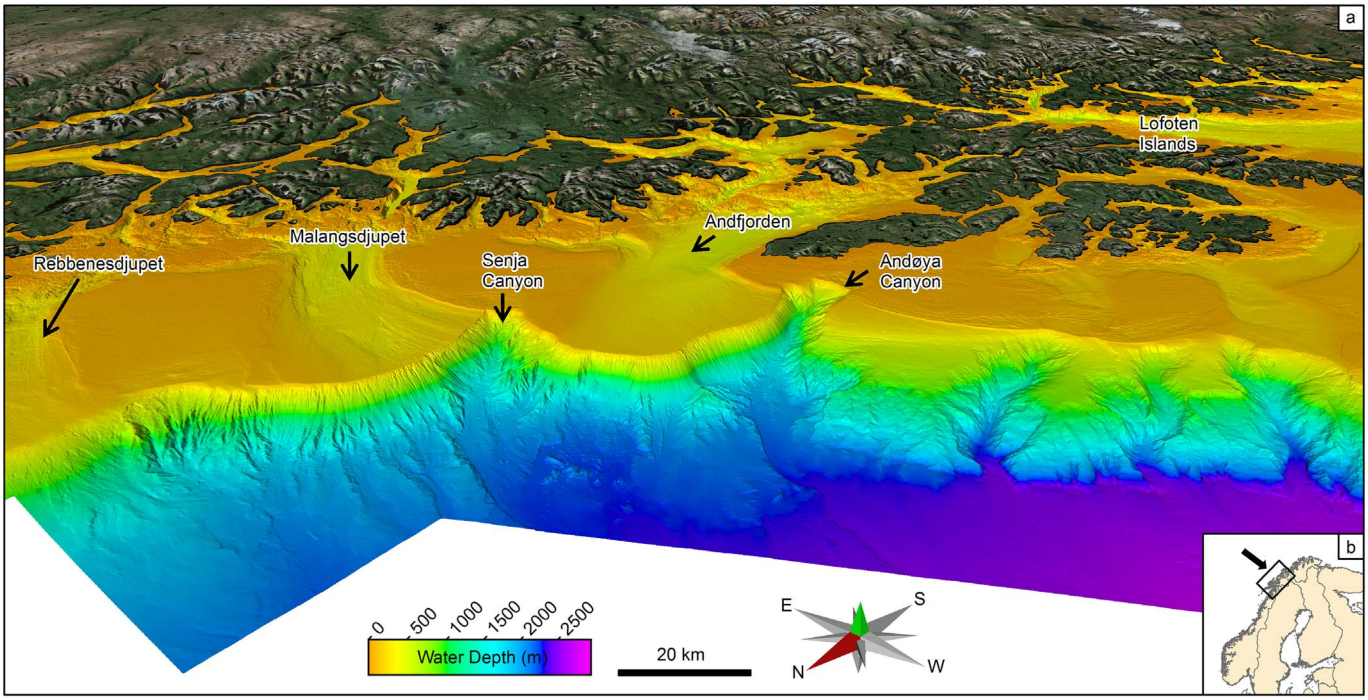

Fig. 8. a) 3D image showing canyons incised into the Norwegian shelf north of the Lofoten Islands. The morphology here is markedly different to other parts of the Norwegian shelf. Note how the trough-mouth fans are separated by canyon features along most of this part of the margin. The bathymetric data are provided from the MAREANO multibeam data set (available at www.mareano.no) collected by the Norwegian Hydrographic Service. Areas discussed in the text such as the Andøya Canyon are annotated. b) location on the Norwegian margin of panel (a).

from the Tampen area (34/7-1, 34/4-7, and 34/8-1 on Fig. 1a) show that the Naust Formation equivalent in the northern North Sea is dominated by silt and sand-rich diamict with occasional crystalline pebbles sourced from the Norwegian mainland and was deposited in glacial environments (Eidvin et al., 1998; Eidvin and Rundberg, 2001). Foraminiferal assemblages from cored material in the Tampen area, and correlation to planktonic fauna from the Vøring Plateau, suggest that the cored material is older than 2.4 Ma (Eidvin et al., 1998, 2000).

Early work suggested that the initial buildout of the Naust Formation could be attributed to fluviomarine and gravity-driven processes without the expansion of ice onto the continental shelf (Bugge et al., 2004). However, as pointed out by Ottesen et al. (2009), the shelf break at the beginning of deposition of the Naust Formation was the delta front of the Molo Formation. Thus, Early Pleistocene glaciers had a shorter distance to travel to reach the shelf edge compared to the Late Pleistocene. Regional 2D seismic profiles across the Naust Formation show that during N-unit time (2.8–1.5 Ma) the shelf prograded up to ~80 km (Fig. 3). Whilst the early period of glaciation on the northern part of the margin appears to have been dominated by a glaciomarine and glaciofluvial phase (Laberg et al., 2010, 2011; Rydningen et al., 2016), the identification of buried landforms within the Naust Formation offshore mid-Norway shows that ice extended onto the continental shelf during the deposition of the N-unit (Ottesen et al., 2009; Montelli et al., 2017). This suggests that ice sheets drove margin progradation offshore mid-Norway as glaciers delivered material directly to the shelf and the shelf edge. Consequently, margin progradation offshore mid-Norway can be related to the increased availability of sediments over the last 2.8 Ma due to increased erosion and uplift in response to the onset of major glaciations over Northwest Europe (Henriksen and Vorren, 1996; Stoker et al., 2005; Ottesen et al., 2009).

3.3. Margin architecture

The evolution of high-latitude passive margins is often dominated by the degree to which ice extended across the continental shelf and the rate of sediment delivery to the margin (Dowdeswell et al., 1996; Taylor et al., 2000). Trough mouth fans (TMFs) form basinward of cross-shelf troughs that were once occupied by ice streams that delivered large volumes of sediment to the shelf edge (Vorren and Laberg, 1997; Vorren et al., 1998). This material was frequently remobilised as glaciogenic debris flows (GDFs) down the continental slope (Vorren and Laberg, 1997; Ó Cofaigh et al., 2003). GDFs are formed from gravity-driven processes in ice-marginal areas (Vorren et al., 1989; Dahlgren et al., 2005) and are a unique mid- to high-latitude type of debris flow that share characteristics with fluidised, liquefied, and classical debris flow types. The abundance of immature clasts of widely varying lithologies is typical of GDFs compared with regular debris flows deposits (Taylor et al., 2002). As diamict-dominated material is delivered to the margin, a convex, seaward-outbuilding, fan-shaped feature develops (Vorren and Laberg, 1997). Sediment accumulation beyond the shelf edge is not exclusively the result of GDFs but also from contour currents (van Weering et al., 2008) and the settling of meltwater plumes (Taylor et al., 2002). Thus, TMF development reflects the combination of shelf morphology and glacial and oceanographic processes.

TMFs are generally restricted to the shallowest of slopes (< 4°), and on margins with steeper slopes or lower sediment fluxes, these areas are often characterised by slope failure or submarine channel and gully complexes caused by erosive turbidity currents rather than TMFs (Talling et al., 2002; Ó Cofaigh et al., 2004; Wilken and Mienert, 2006; Piper and Normark, 2009). The transition between TMF and channel complex development is influenced by grain size distribution, the flux of sediment-laden meltwater (Ó Cofaigh et al., 2003; Wilken and Mienert, 2006), and the inherited geologic setting (e.g. steep slopes and sediment bypass) (Batchelor and Dowdeswell, 2014). A large part of the Norwegian margin is characterised by TMFs, from the North Sea Fan at the end of the Norwegian Channel (Nygård et al., 2005), to the small TMFs at the end of the Rebbenesdjupet, Malangsdjupet, and Andfjorden cross-shelf troughs (Fig. 8) offshore Troms (Rydningen et al., 2013). Although the mid-Norwegian margin is generally classified as a prograding wedge (Dahlgren et al., 2005), it is a product of similar depositional processes operating on the TMFs elsewhere along the margin.

3.4. Submarine canyons

The shelf edge to the north of the main sedimentary wedge on the Norwegian margin is characterised by several TMFs with intra-TMF areas characterised by canyons such as the Andøya and Senja Canyons (Laberg et al., 2006; Rydningen et al., 2015) (Fig. 8). Up to 700 m of interbedded turbidites and hemipelagic sediments have been deposited at the mouths of these canyons (Rise et al., 2013), demonstrating the large flux of material from the shelf to the basin in these areas. The canyons off northern Norway (Fig. 8) owe part of their development to the glacial evolution of adjacent TMFs. The Andfjorden, Malangsjupet, and Rebbenesdjupet TMFs have been affected by both large- and small-scale submarine landslides in glaciomarine sediments and/or contourites. Erosive turbidity currents, which originated from debris flows on the steep upper slopes of the TMFs, have helped to widen and deepen canyons in the intra-TMF areas as sediment was transported down the slope (Rydningen et al., 2015).

The Andøya Canyon (Fig. 8) has been excavated from the "bottom-up" as an initial slope failure has allowed the scar to be further cut by failure of the headwall and sidewalls as well as turbidity current erosion by materials delivered by ocean and ice sheet processes (Laberg et al., 2007). The highest frequency of turbidite deposition occurred when the local Andfjorden Ice Stream reached the shelf break near to the canyon head and delivered a constant supply of sediments (Amundsen et al., 2015). Increased northward currents through the last deglaciation have later led to winnowing and failure of sediments in the southern part of the canyon, leading to further incision (Amundsen et al., 2015). A major submarine channel system connects upslope to the Andøya Canyon, with the majority of the sandy material deposited by turbidity currents that pre-date the Holocene (Ó Cofaigh et al., 2006). In the Holocene succession, very few slide deposits are evident and recent activity has seen turbidity currents flush and erode the thalweg of the canyon (Laberg et al., 2007; Amundsen et al., 2015). Taken together, when ice sheets were at their greatest extent, the direct delivery of material to the margin led to gully and slide development on the upper slope. Remobilised material was then funnelled through the intra-TMF areas where erosive turbidity currents aided canyon formation. As ice sheet margins became more distal, oceanographic processes drove further turbidity current formation, but at a reduced frequency.

3.5. Along-slope processes

The Atlantic margin of Norway has been subjected to along-slope processes through much of the late Cenozoic (Fig. 1a). From the Miocene onwards, a large increase in the growth of contourite drifts occurred (Laberg et al., 2001) and it is thought that this reflects increased strength in ocean currents in the North Atlantic caused by the northward penetration of warm surface waters (Galloway, 2002; Laberg et al., 2005b) (see section 4.1).

In the sedimentary succession preserved on the Vøring Plateau (Fig. 1a), seismic profiles tied to borehole data (Fig. 9a) show that ocean currents have influenced sedimentation patterns through the Paleogene and into the late Miocene (Laberg et al., 2005a). After this, the Vøring Plateau provided a reduced topographic control on ocean currents crossing the area, resulting in a switch from current-controlled, to hemipelagic sedimentation from the early Pliocene (Laberg et al., 2005a). Elsewhere, mounded contourites (Fig. 9b) were formed during the Miocene when structural features such as the HHA and the Nyk High (Fig. 1a) provided a topographic control that constrained ocean currents sweeping the area (Hjelstuen et al., 2004a; Laberg et al., 2005b).

Offshore the Lofoten Islands on the northern part of the margin, two elongated drift deposits are present; the Lofoten (Fig. 10) and Vesterålen Drifts (Laberg et al., 1999, 2001, 2005b). The drifts are comprised of sediments accumulated from the mid Miocene to present (i.e. the Kai and Naust Formations) and they both overlie an intra-Miocene unconformity (Blystad et al., 1995), meaning the currents intensified during the Miocene (Laberg et al., 1999, 2001, 2005b). During the last glacial-interglacial cycle, climate changes influenced deposition as the highest rates of sedimentation were recorded during maximum ice sheet conditions (Laberg and Vorren, 2004).

By the late Pliocene and Pleistocene, subsidence of the Vøring Plateau meant that it provided less topographic influence on ocean currents, which were now likely shifted eastward toward the continental slope and shelf (Laberg et al., 2005a, 2005b). This period was characterised by glacial-interglacial cycles and different types of glaciogenic sedimentation occurred along the margin (Dahlgren et al., 2005; Nygård et al., 2005; Ottesen et al., 2009; Rydningen et al., 2016). Contourites were deposited in two different settings. In the first, mounded drifts such as the Nyk Drift are intercalated between different units of GDFs deposited during the penultimate and the last glacial cycles (Laberg et al., 2001). Part of this drift deposit has also been significantly reworked by the later Nyk and Trænadjupet Slides discussed in section 3.6 (Laberg et al., 2003; Lindberg et al., 2004). In the second setting, drift growth has occurred within the slide scars formed by submarine landslides (Dahlgren et al., 2002; Rise et al., 2006b). Except for during glacial maxima, when glaciogenic deposition dominated (Laberg et al., 2005b), the winnowing of sediments and infilling of the slide scars operated through to the Holocene when present-day circulation patterns of the North Atlantic were established (Vorren et al., 1984; Laberg et al., 2003; Haflidason et al., 2005). Similar deposits have been observed within the stratigraphy of the Storegga Slide scar that pre-date the Holocene (Fig. 11), thus suggesting that sedimentation similar to the present also occurred in the past (Sejrup et al., 2004, 2005).

3.6. Mega-slides

Except for submarine slide scars (e.g. Fig. 11), the morphology of the margin has been largely untouched since ice retreated onshore at ~14–13 ka (Hughes et al., 2016; Stroeven et al., 2016) and clastic sediments became trapped in the Norwegian fjords (Vorren et al., 1989; Rise et al., 2006a; Hjelstuen et al., 2009). Thus, large areas of the shelf have experienced little deposition, with bottom currents winnowing sediments throughout the Holocene (Ottesen et al., 2009). A number of slides have been identified along the margin (Kenyon, 1987; Vorren et al., 1998; Leynaud et al., 2009) and these vary in size from the large Storegga Slide (Fig. 11), which covers an area of 95,000 km^2 (Haflidason et al., 2005; Solheim et al., 2005), to the smaller Andøya Slide (Fig. 1a) which covers an area < 10,000 km^2 and comprises multiple slides (Laberg et al., 2005c; Rise et al., 2013). These slides and their associated sedimentary processes are important for understanding the longer-term depositional history of the deep-water basins adjacent to the Norwegian margin as they reflect wider tectonic, oceanographic, and climatic changes.

The main Storegga Slide (Fig. 11) was a retrogressive event that mobilised a sediment volume of up to 2400–3200 km^3 at ~8.2 ka BP (Haflidason et al., 2004, 2005). This date corresponds with tsunami deposits along the coastlines of Greenland and Northwest Europe (Grauert et al., 2001; Bondevik et al., 2003; Tooley and Smith, 2005; Wagner et al., 2007; Vasskog et al., 2013). Since the last major slide at ~8.2 ka BP, only a few minor slumps have occurred at 5.7 ka and 2.8–2.2 ka BP (Haflidason et al., 2005). The Holocene and pre-Holocene mega-slides of the area have common characteristics that can be related to the cyclicity of glacial-interglacial sedimentation (Bryn et al., 2005). During glacial maxima, rapid deposition of diamicts occurred at the shelf edge and GDFs on the slope were associated with the North Sea Fan deposition basinward of the Norwegian Channel Ice Stream (Hjelstuen et al., 2004b; Sejrup et al., 2004) (Fig. 11). During interglacials, finer-grained hemipelagic deposits and contourites prevailed in the basin and slope areas (Sejrup et al., 2004; Micallef et al., 2009). In the early stages of compaction the hemipelagic sediments were more

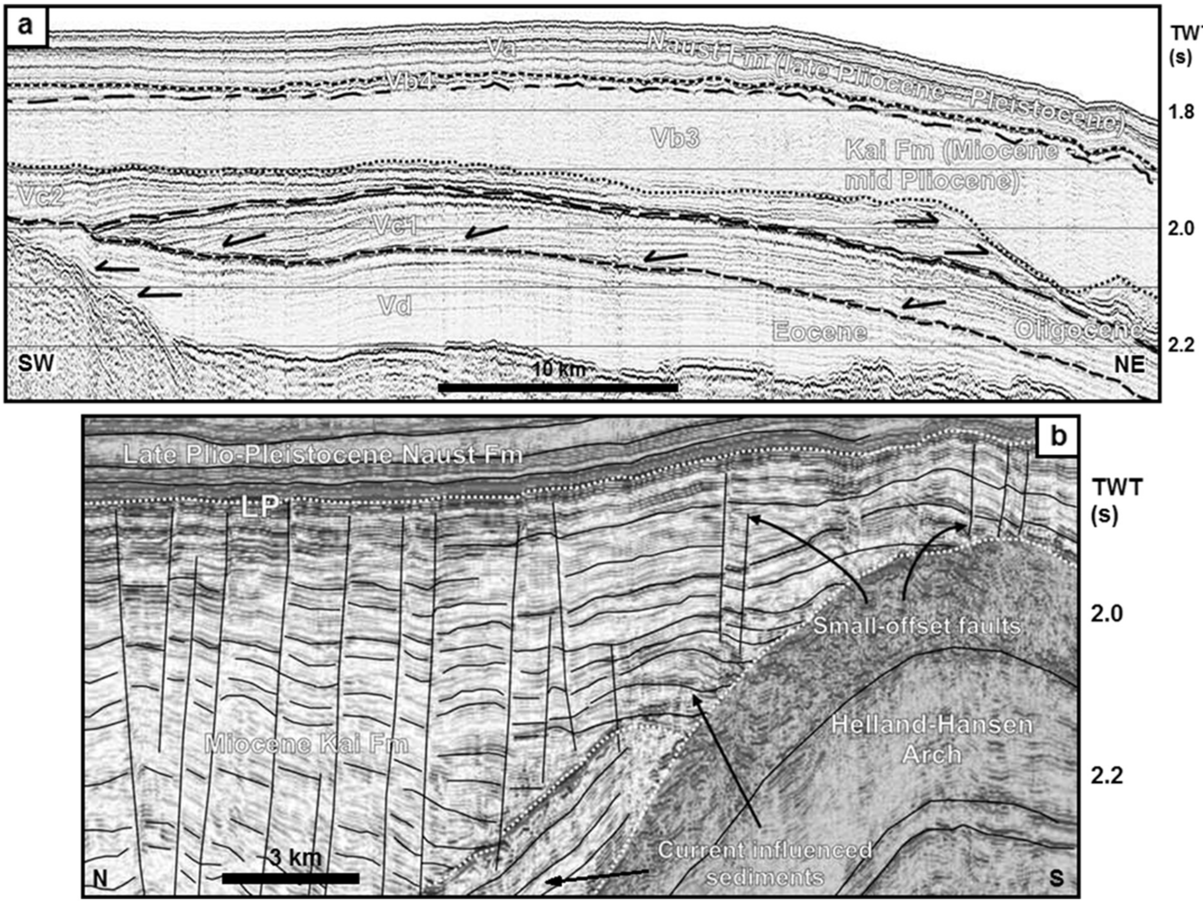

Fig. 9. a) Seismic profile from the northeastern part of the Vøring Plateau showing the onlapping and downlapping within the Vd, Vc1, Vc2, and Vb3 seismic units. Deposition is thought to have been controlled by ocean currents through much of the Cenozoic until in the youngest stratigraphy (units Vb4 and Va), where hemipelagic sedimentation dominates as the ocean currents in the area have migrated. Figure modified from Laberg et al. (2005a, 2005b). Location of profile shown on Fig. 1a. b) Ocean current controlled deposition onto the western flank of the structural high, the Helland-Hansen Arch on the southern part of the Vøring Plateau. LP refers to the base of the late Pliocene and Pleistocene glaciogenic Naust Formation. Figure modified from Hjelstuen et al. (2004a) and Laberg et al. (2005b). For location see Fig. 1a.

permeable than glacial sediments (Berg et al., 2005) and pore pressures increased in the hemipelagic deposits during the glacials due to rapid sedimentation as well as fluid flow from underlying oozes (Solheim et al., 2005). Hydrate dissociation prior to the main Storegga event may have weakened layers within the sediment (Mienert et al., 2005) that then failed due to seismicity triggered during ice sheet retreat and isostatic rebound (Kvalstad et al., 2005; Solheim et al., 2005). The area is particularly prone to slope failure and generation of glide planes and is thought to have occurred through a number of different glacial periods (Solheim et al., 2005).

The Trænadjupet Slide was initiated during the mid-Holocene at ~4 ^{14}C ka BP in an area southwest of the Lofoten Islands (Fig. 1a). The slide affected an area of 14,100 km^2 and like the Storegga Slide region, high sedimentation rates dominated during peak glacials and hemipelagic deposits and contourites prevailed during interglacials. The hemipelagites and contourites later failed due to loading by glaciogenic

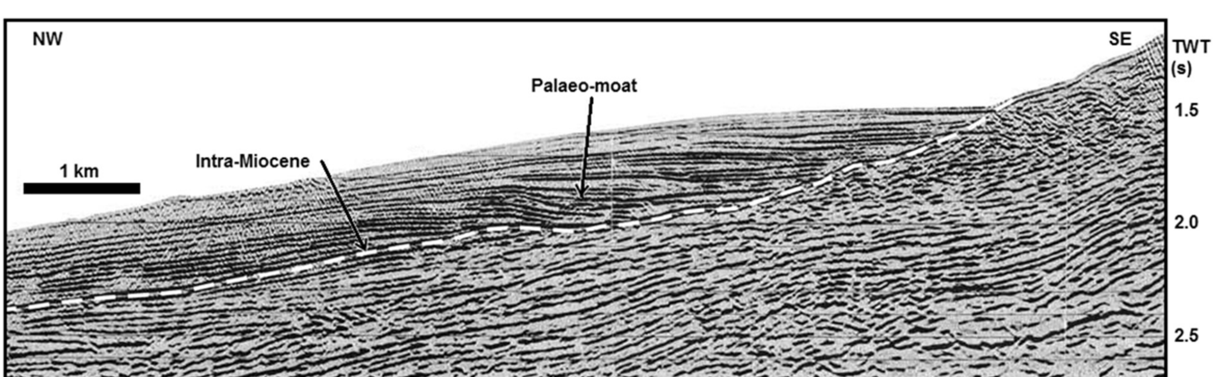

Fig. 10. Seismic profile across the Lofoten Drift on the northern part of the margin. The base of the drift is thought to be Miocene in age, although a precise date is lacking. The acoustic character shows that the internal reflections are layered with a parallel to slightly diverged pattern. A palaeo-moat shows migration of the drift deposit through time. Figure modified from Laberg et al. (2001, 2005b). Location of the profile is shown on Fig. 1a.

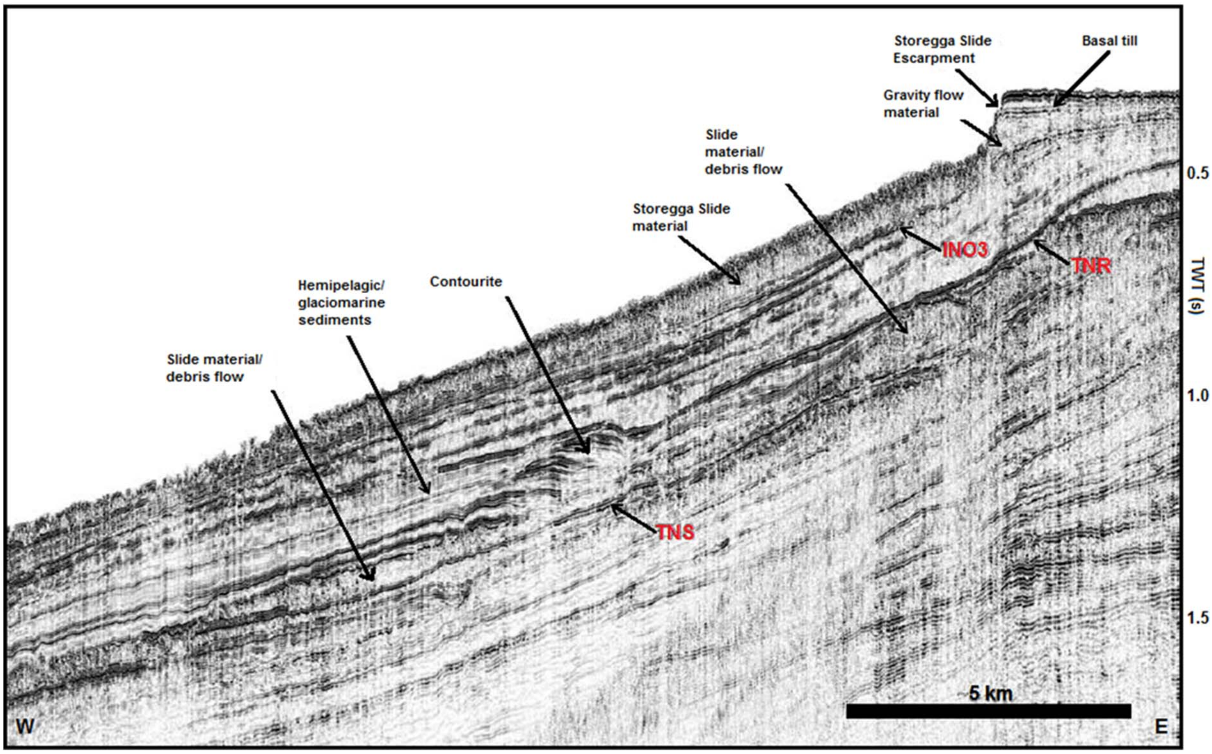

Fig. 11. Seismic architecture of the Storegga Slide region on the southern part of the margin. Boundaries of the different seismic units are as follows: the Storegga glide plane (INO3), and top of the Naust R (TNR) and S (TNS) units. These relate to one of the older stratigraphic systems shown in Fig. 6. On this part of the margin, contourite and hemipelagic deposits are deposited on top of glaciogenic debris flows and can act as fault planes when they are buried during periods of intense deposition. Figure modified from Sejrup et al. (2004) and Laberg et al. (2005b). For location, see Fig. 1a.

sediments and were possibly triggered by Holocene seismicity (Laberg and Vorren, 2000; Laberg et al., 2002, 2006).

Piston core studies have shown the frequency of submarine landslides in the Lofoten Basin (Laberg et al., 2006). Five intervals of turbidite deposition are separated by long periods of hemipelagic deposition and the first interval of turbidites has been correlated with the Trænadjupet Slide and the second interval with the Nyk Slide at ~16 ^{14}C ka (Laberg et al., 2006) (Fig. 1a). The Nyk Slide covers at least 2200 km^2 with parts of the deposit being removed by the later Trænadjupet Slide (Lindberg et al., 2004). Gravity cores have shown that the slide was preceded and succeeded by the deposition of GDFs, indicating the presence of ice at the shelf edge (Lindberg et al., 2004). The glide planes for both the Trænadjupet and Nyk Slides are located within contourites with high water content that helped promote failure (Baeten et al., 2014). Laberg et al. (2006) tentatively correlated the three older turbidite intervals to MIS 6, 10, and 12, when ice advanced to the shelf edge; thus underlining the key influence of rapid deposition of glacial sediments and over-pressuring of hemipelagic deposits in developing the initial conditions that promote slope failure.

Several buried slides along the margin have also been identified. These include the Møre and Tampen Slides of the North Sea Fan (Fig. 1a) that are related to rapid sedimentation from the Norwegian Channel Ice Stream (King et al., 1996; Nygård et al., 2005). Rise et al. (2006b) investigated the region southeast of the Vøring Plateau where the Sklinnadjupet Slide removed sediments from the Naust Formation S-unit. A palaeo-trough extends across the shelf on Haltenbanken and Trænabanken (Fig. 1a) leading to a large depocentre. Like the other slides described above, rapid sedimentation directly onto the shelf edge and slope contributed to promoting high pore pressures and subsequent slope failure during an earthquake (Rise et al., 2006b). The slide also eroded part of the HHA anticline as diatomaceous oozes were mobilised and material was deposited down-slope as levee ridges and mounds above slide deposits from the earliest phase of slope failure (Rise et al., 2006b).

3.7. Margin subsidence

3.7.1. Neogene controversy

The long-term, large-scale origin of the Scandinavian high topography has been the cause of controversy in the literature as researchers largely fall into two schools of thought. The first paradigm suggests that isostasy, climate, and erosion have been the key factors leading to the development of the western Scandinavian topography. Repeated climate deterioration and erosion of the mountains by fluvial, glacial, and periglacial processes led to exhumation and dissection of a low-elevation peneplain since the Caledonide Orogeny (490–390 Ma). This resulted in a compensating flexural isostatic response that continued through the late Cenozoic and does not require active tectonism during the Neogene to achieve uplift (Clausen et al., 1999; Huuse, 2002; Nielsen et al., 2009; Gołędowski et al., 2013). The key argument of the second school is that the basement surface uplift of the western Scandinavian led to the denudation of the topography and construction of Phanerozoic peneplains that were later buried before being re-exposed during three phases of Cenozoic tectonism (Japsen and Chalmers, 2000; Japsen et al., 2007; Lidmar-Bergström et al., 2013). These authors suggest that continuous uplift cannot be assumed for the region and the isostatic response of the mountain chain to fluvial and glacial erosion, as preferred by the first paradigm, has only served to amplify the underlying tectonic uplift of the mountains. Between the two end members a more detailed argument has been made for a combination of differential neotectonic movements but also honouring the obvious effects of glacial erosion unloading and sedimentary loading (Redfield et al., 2005). However, the fundamental controls on mountain growth in the region remain enigmatic and different paradigms for their evolution remain. Resolving this debate is not currently possible and we refer those interested to the papers referenced above as a starting point. Here, we focus on the Plio-Pleistocene, where the topographic response to denudation and sedimentation has been largely driven by glacial-interglacial cycles.

3.7.2. Plio-Pleistocene subsidence

The main wedge of glaciogenic material offshore mid-Norway is shown to have preserved a number of the palaeo-shelves (Fig. 3) (Ottesen et al., 2009). Similar patterns of increasingly buried palaeo-shelf edges are also observed for the North Sea Fan to the south of the mid-Norwegian margin (King et al., 1996; Sejrup et al., 2003). The greater preservation of the palaeo-shelf edge is related to erosion of the inner shelf and rapid deposition of sediments when ice was grounded on the shelf. This generates subsidence of the outer shelf and increases accommodation available for sediment accumulation during subsequent glacials (Powell and Cooper, 2002). Coeval with subsidence of the outer shelf, the inner shelf and the Norwegian mainland were also likely uplifted in response to denudation and lower average sea level during the Pleistocene (Doré, 1992; Huuse, 2002; Dowdeswell et al., 2010; Ottesen et al., 2012). Bedrock lowering of ~400–520 m through the glacial period has been estimated from seismic data, numerical modelling, and apatite fission track analysis (Dowdeswell et al., 2010; Medvedev and Hartz, 2015) and much of this eroded material is deposited on the middle and outer shelf (Ottesen et al., 2012).

Subsidence along the margin varies significantly and is coincident with variability in depositional patterns (Dahlgren et al., 2005; Hjelstuen et al., 2005). Rates of subsidence are generally highest on the outer shelf and slope of the mid-Norwegian margin (Dahlgren et al., 2005). Estimates of subsidence rates along several parts of the margin have been based on the identification of glacial erosion surfaces, maximum depths for glacial geomorphological features and deltaic deposits. These show subsidence estimates of 0.1–1.2 m ka^{-1} since ~350 ka (King et al., 1996; Lebesbye and Vorren, 1996; Solheim et al., 1996; Dahlgren et al., 2002; Sejrup et al., 2004).

3.7.3. Glacial isostasy

On shorter timescales, as ice sheets grow and decay through different glacial-interglacial cycles, the Earth's mantle and crust respond to changes in ice loading. During glacial periods, the weight of the ice sheet, which could reach thicknesses up to 4.5 km thick in Northwest Europe (Lambeck et al., 2006), would depress the Earth's crust by several hundreds of metres. Indeed, concomitant with a glacio-eustatic drop in sea level, the forebulge generated in the Barents Sea from crustal depression beneath Fennoscandia from ice loading reduced water depths by up to 200 m, which then aided further ice sheet growth (Howell et al., 2000). As the ice cover over Fennoscandia gradually retreated from the shelf to its current high altitude locations during the last glacial cycle, the reduction of this load allowed the Earth's crust to rebound. This rebound, termed glacial isostatic adjustment, is still observable today due to the time-dependent viscoelastic response of the Earth's mantle to the release in stress (Lambeck et al., 1998b; Zhao et al., 2012).

Northwest Europe has been a key area for researching glacial isostatic adjustment as a large number of geological and geophysical observations have been recorded to aid the spatio-temporal reconstruction of isostatic movements since the LGM (Steffen and Wu, 2011). Glacial to interglacial palaeo-strandlines are sparse for the earliest parts of the last glacial cycle, particularly in Fennoscandia (Lambeck et al., 2010). However, records of post-LGM palaeo-strandlines, tidal gauges, and uplifted marine sediments have been observed across coastal Northwest Europe (Lambeck et al., 1998a, 2010; Steffen and Wu, 2011). These records have been used to develop local relative sea level curves through the Holocene for a large number of study sites across Fennoscandia (Schmitt et al., 2009; Romundset et al., 2010) and the southern Baltic Sea (Bungenstock and Schäfer, 2009; Gelumbauskaitė, 2009; Hijma and Cohen, 2010). Crustal deformations and changes in the Earth's gravity field have also been measured using global positioning systems (GPS) (Johansson et al., 2002; Lidberg et al., 2007) and satellite data such as from the Gravity Recovery and Climate Experiment (GRACE) (Steffen et al., 2008, 2009), respectively.

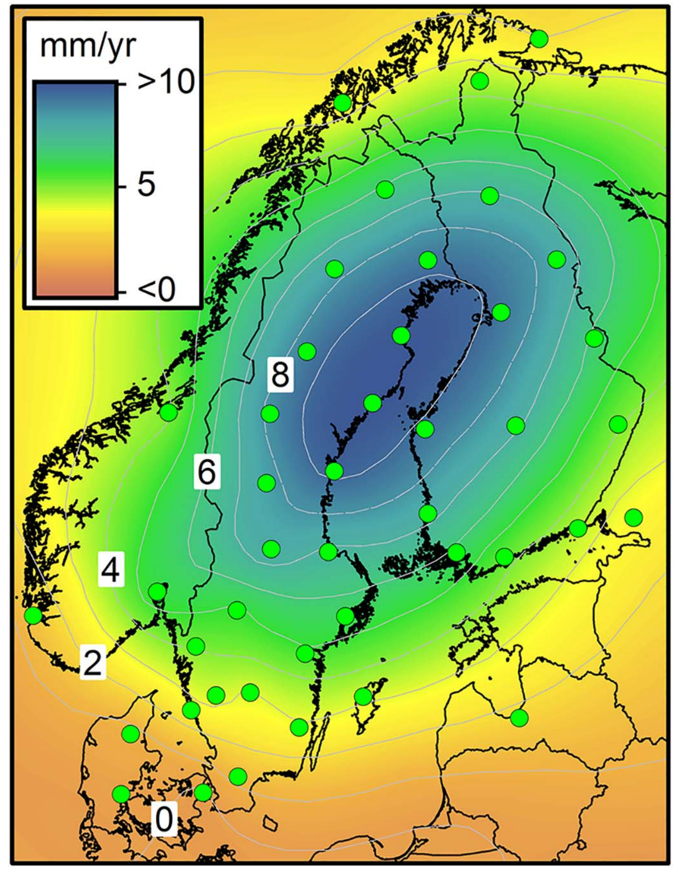

Fig. 12. Compilation figure showing the contemporary isostatic rebound of Fennoscandia. Green circles mark locations where uplift is being recorded on a continuous basis by GPS measurements. Figure compiled from Rosentau et al. (2012) and Johansson et al. (2002).

These methods can be combined with absolute gravimetry collected from field campaigns and numerical modelling techniques to obtain a data-based understanding of past and contemporary deformation and mass variation across Fennoscandia (Wu et al., 1999; Müller et al., 2012; Nielsen et al., 2014). Using these different methods, contemporary uplift rates across Fennoscandia are in the region of ~10 mm yr^{-1}, whilst the total uplift since the end of the last glaciation may be up to several hundred metres, particularly near to the centre of the Baltic Ice Dome (Steffen et al., 2014) (Fig. 12). This rate varies across the region with maximum uplift rates of 11–20 mm yr^{-1} recorded in parts of the Gulf of Bothnia (Johansson et al., 2002; Steffen et al., 2008) and negative rates beyond the former ice margin (Fjeldskaar et al., 2000).

These studies of isostatic adjustment have been used to suggest that the mountains were more glacio-isostatically depressed than previously thought (Kolstrup and Olsen, 2012). If such findings are later validated, changes in subsidence and uplift in response to glacial isostasy will be important for understanding how glacial isostatic adjustment impacts buried hydrocarbons in frontier glaciated basins. Estimates suggest that up to 30% of the total hydrocarbon volume generated in high-latitude petroleum systems could be lost during isostatically-driven changes in burial depth of potential plays, and also concomitant tilting and exhumation of hydrocarbons traps (Kjemperud and Fjeldskaar, 1992; Tasianas et al., 2016; Zieba and Grøver, 2016). Understanding glacial isostatic adjustment will become increasingly important, not just because of the impact on buried hydrocarbons, but also for coastal communities and whether the rate of isostatic rebound (or subsidence) outpaces the rate of global sea level rise (Hill et al., 2010; Grinsted et al., 2015).

4. Oceanography

4.1. Palaeo-oceanography of the Norwegian Sea

During the early Cenozoic, when rifting of the region was in its infancy (Fig. 2), ocean circulation within the proto-NGS was heavily restricted due to the limited interaction between the Atlantic and Arctic Oceans (Thiede et al., 1996). As the North Atlantic widened in the Oligocene and Miocene (Brozena et al., 2003), the Greenland-Scotland Ridge subsided in response to lithospheric cooling (Poore et al., 2006), and the Fram Strait opened (Engen et al., 2008). An early to middle Miocene deepwater connection was then established with the Arctic Ocean and circulation patterns akin to today were developed (Wright and Miller, 1996; Laberg et al., 2005b; Jakobsson et al., 2007). It should also be noted that an Eocene to early Oligocene onset for NADW production, based on drift deposits offshore the Faroe Islands, has been proposed (Davies et al., 2001; Hohbein et al., 2015). However, this model is not widely accepted and requires further validation by palaeo-oceanographic proxies.

A major global climate shift occurred during the Pliocene as repeated glacial-interglacial cycles became established (Lisiecki and Raymo, 2005). The Northern Hemisphere ice sheets first expanded at ~2.8 Ma when IRD was measured in the Atlantic and Pacific Oceans (Rea and Schrader, 1985; Jansen and Sjøholm, 1991; Raymo, 1994). Tectonic, orbital, and climatic mechanisms are thought to have caused this shift and are outlined in more detail by Lisiecki and Raymo (2007). Of particular interest here are oceanographic changes offshore Norway during this period. The NAC is a crucial part of the global thermohaline circulation as it brings heat and moisture to the high latitudes of the North Atlantic (Lunt et al., 2008). As the NAC splits before continuing its flow through the Norwegian Sea, its Plio-Pleistocene evolution has an important influence on climate and associated depositional environments along the Norwegian margin. Any fluctuations in the strength of the NAC, and the location of NADW production, will influence the Norwegian Current and ice sheet growth and decay (Broecker, 1991; Clark et al., 2002). The record of glaciations offshore Norway indicates three main palaeoceanographic regimes (Henrich, 1989), which are detailed in sections 4.2–4.4; (i) 41 kyr glacial cycles during the late Pliocene and Early Pleistocene, (ii) the Mid-Pleistocene Transition (MPT) from 1.2–0.7 Ma when these cycles were gradually replaced by higher amplitude 100 kyr cycles, and (iii) the Middle Pleistocene to Holocene which was dominated by 100 kyr cycles (Fig. 7).

4.2. Late Pliocene and Early Pleistocene

Although it is thought that global ocean circulation patterns reorganised after the closure of the Isthmus of Panama (Haug and Tiedemann, 1997), it is uncertain how the ocean-climate system changed at the intensification of the Northern Hemisphere glaciations. It has been proposed that the NAC might have shifted away from the Norwegian margin to lower latitudes, causing a southward shift of the Arctic Front (Hennissen et al., 2014). However, there is also evidence that the NAC showed little variation and extended northward into the Norwegian Sea (Friedrich et al., 2013). The evolution of the NGS currents to something similar to today is important for NADW production and a slowdown in northward heat transport was possibly the precursor for late Pliocene glaciation of Greenland (Naafs et al., 2010; De Schepper et al., 2015). As the 41 kyr Northern Hemisphere glacial-interglacial cycles became the climatic norm during the Early Pleistocene, the North Atlantic circulation appears to have changed little compared to the temperature changes (Raymo et al., 2004). An exception to this is observed in the Norwegian Sea as warmer Atlantic waters episodically penetrated along the eastern margin (Henrich and Baumann, 1994; Henrich et al., 2002). These intrusions have been inferred from biogenic calcareous records and the presence of warm adapted species such as those found at ODP Site 644 (Fig. 1a) (Baumann et al., 1996). IRD fluctuations in the North Atlantic suggest that after the first major glaciation of Fennoscandia at ~2.8 Ma, decreased carbonate production occurred as perennial or permanent sea ice cover in the Norwegian Sea helped reduce ocean overturning (Henrich, 1989; Huber et al., 2000a) (Fig. 7). Prior to this (and like in the present North Atlantic (Huber et al., 2000b)), carbonate had been well-preserved in the Pliocene and provides supporting evidence for NADW production in the NGS prior to major glaciation (Henrich et al., 2002).

4.3. Mid-Pleistocene Transition (MPT)

A profound climatic shift started at ~1.2 Ma (Fig. 7) when the dominant periodicity of the glacial cycles began to shift from 41 kyr to 100 kyr cycles. It has been suggested that the growth of the Antarctic Ice Sheet (Elderfield et al., 2012) and reduced strength of the AMOC during MIS 23 and a marked deterioration of the climate during MIS 22 may have been responsible for this shift (Hernández-Almeida et al., 2013; Pena and Goldstein, 2014). This weakening of the AMOC predated the main phase of ice sheet growth (Pena and Goldstein, 2014) and helped to stabilise the 100 kyr cycles, create larger ice sheets, and develop an overall greater contrast between the glacial-interglacial cycles than in the Early Pleistocene (Clark et al., 2006; Huybers, 2006; Hönisch et al., 2009).

In the Norwegian Sea, changes at the MPT are reflected by a marked increase in IRD after ~1.1 Ma (Krissek, 1989; Jansen et al., 2000) (Fig. 7). The increased duration of glacial periodicity allowed ice sheets to grow beyond the coastline and onto the continental shelf west of Norway. The Norwegian Channel Ice Stream is also thought to have been established at this time (Sejrup et al., 1995) and may have contributed to the IRD supply to the Norwegian Sea. Palaeotemperatures inferred from *Neogloboquadrina pachyderma* (sin.) offshore Norway indicate an increased contrast between glacial-interglacial environments and an increased meridionality in circulation strengthening in the NGS (Henrich, 1989; Henrich and Baumann, 1994; Huber et al., 2000b). The NAC and its northward extensions through the Norwegian Sea were well-developed during the peak interglacials, whilst during the glacial cycles the surface waters were dominated by sea ice cover, iceberg drift, and freshwater melting (Henrich and Baumann, 1994; Baumann et al., 1996). During deglaciation, the reduced magnitude of NADW in the NGS was occasionally interrupted by ice sheet collapse and the development of an extensive meltwater lid that restricted heat advection through the Norwegian Sea (Henrich et al., 2002).

4.4. Middle to Late Pleistocene and Holocene

The Middle and Late Pleistocene were dominated by 100 kyr glacial cycles that led to the growth of huge continental ice sheets, as is reflected in carbonate and IRD depositional patterns (Henrich, 1989) (Fig. 7). The mid-Brunhes Event is the most pronounced climatic shift through the Middle and Late Pleistocene. In the Nordic Seas (above 56°N) this is characterised by colder interglacial periods from MIS 19-13 when compared to those during MIS 11-1 (Candy and McClymont, 2013). This is thought to relate to the position of the Polar Front and its extension to latitudes below 61°N during some interglacials prior to MIS 12 (Wright and Flower, 2002).

Generally it has been suggested that NADW occurred more or less continuously through the Middle and Late Pleistocene glacial-interglacial cycles in the NGS with variations in the degree, depth, and location of overturning and the Norwegian Current (Seidov et al., 1996; Levine and Bigg, 2008; Newton et al., 2016). Peak carbonate dissolution since ~0.6 Ma indicates reduced NADW production (Huber et al., 2000a) due to a reduced Norwegian Current, sea ice cover, periods of high meltwater flux (Kellogg, 1977), and iceberg armadas that blocked surface circulation near the Greenland-Scotland Ridge (Henrich, 1989). In general, very little is known about the ocean dynamics and timescales of the oldest glacial-interglacial cycles as most work has focussed on high-resolution records that capture the last 100 kyr cycle.

During the Eemian interglacial (MIS 5e), sea levels were 5–9 m higher (Dutton et al., 2009, 2015) and Arctic temperatures were 2–4 °C higher than today (van de Berg et al., 2011). The climate in the NGS during the Eemian is thought to have been less stable with water masses distributed differently to the current interglacial (Fronval and Jansen, 1997). The NAC extended far into the NGS until ~115 ka when a drop in sea surface temperature indicates southward movement of the Polar Front and the NAC as ice sheet growth began (Müller and Kukla, 2004). Through the Weichselian (MIS 5d-2), a large number of geological and numerical modelling studies have investigated the evolution of the NAC, the Norwegian Current, and more broadly the AMOC. Deepwater production occurred almost continuously during the Weichselian and three different modes of NADW are thought to have existed; the glacial, the Heinrich, and the modern modes of circulation (Bigg et al., 2000; Rahmstorf, 2002; Lynch-Stieglitz et al., 2007). As the circulation mode changed there were rapid changes in ocean overturning and heat flux through the Norwegian Sea. This in turn had a major impact on the climate of Northwest Europe and ice sheet mass balance (Fronval and Jansen, 1996; Gherardi et al., 2009).

The glacial mode is characterised by reduced overturning with a small amount of seasonal deepwater being produced in the NGS through brine rejection and buoyancy loss during sea ice growth (Vidal et al., 1997, 1998). Most studies investigating the glacial mode have shown similar trends for a much reduced NAC (Negre et al., 2010; Ballarotta et al., 2014) with NADW occurring south of the NGS and the Greenland-Scotland Ridge (Sarnthein et al., 1994; Seidov et al., 1996; Bigg et al., 2012). Overturning was reduced by up to a third (Levine and Bigg, 2008) as extensive sea ice helped to reduce northward advection of heat by up to 50% (Ballarotta et al., 2014). In general, NADW production persisted through the last glacial with significant deviations related to iceberg discharge events, reorganisation of the AMOC, and sudden Greenland temperature changes during Heinrich events (Ślubowska-Woldengren et al., 2008; Böhm et al., 2015).

The Heinrich mode of circulation occurred when large meltwater and iceberg fluxes led to a freshening of surface waters in the NGS and a weakening of the Norwegian Current (Vidal et al., 1997; Chauhan et al., 2016). Modelling and sediment core analyses in the North Atlantic have shown that the meridional overturning was nearly, or completely, eliminated during Heinrich events after the LGM (Gherardi et al., 2009; Bigg et al., 2012; Böhm et al., 2015). The influx of cold freshwater disrupted temperature-salinity profiles in the NGS and affected surface and bottom water circulation (Telesiński et al., 2014). Although the Norwegian Current still propagated northwards along the Norwegian margin, the reduced strength of overturning allowed Antarctic-derived bottom waters to penetrate further into the North Atlantic (Otto-Bliesner et al., 2007; Negre et al., 2010).

The rate of AMOC recovery after Heinrich events varied considerably but was occasionally followed by increased NADW production which were concurrent with warming events (Bitz et al., 2007). Rasmussen et al. (2016) provided a simplified model of the interaction between the glacial and Heinrich modes whereby during warmer periods, heat was pulled into the NGS through active convection. During cooler events convection was limited or stopped completely and heat gradually pushed northward into the NGS (Rasmussen et al., 2016). The rate at which advection of heat occurred would determine how quickly the AMOC would recover after Heinrich events and also the transition from glacial to interglacial ocean circulation patterns.

The modern mode of circulation was first developed ~13.5 ka, although the AMOC remained unstable as circulation in the North Atlantic switched between surface and subsurface advection of heat until ~10 ka when IRD deposition ceased (Bauch et al., 2001). Either side of the Younger Dryas, depletions in epibenthic $\delta^{13}C$ suggest a notable reduction in advection of heat to the North Atlantic (Bauch et al., 2001). In the early Holocene (~7 ka) conditions comparable to today existed, with warm surface waters in the NGS and deep convection and water mass exchange with the Arctic Ocean (Bauch et al., 2001). Increased convection and temperature rise were not coeval across the NGS; the Norwegian margin warmed faster than the central region due to gradual northward and westward expansion of the NAC as the central gyre of the NGS became more established (Telesiński et al., 2015).

As the Holocene thermal maximum ended at ~5 ka, the modern distribution of water masses and east to west surface temperature gradient were developed (Bauch et al., 2001; Meland et al., 2008). Changes in the amount of convection still occurred with weakening of overturning in the Lofoten Basin at ~3 ka, before a gradual increase after ~2 ka as North Atlantic waters gained more influence in the NGS (Telesiński et al., 2015). More recently, the $\delta^{18}O$ record in the Norwegian Sea shows that the Norwegian Current has, over the last 70 years, been at its warmest during the past 600 years (Berstad et al., 2003). It is thought that future warming and freshwater input into the North Atlantic will lead to a weaker barotropic circulation (Nummelin et al., 2015, 2016) and slowing of the AMOC (Boulton et al., 2014). Recent measurements have shown increased freshening and cooling of North Atlantic surface waters and should this continue it could lead to a weakening of the Norwegian Current and thus a weakening of heat advection to Northwest Europe (Rahmstorf et al., 2015).

5. Glacial landforms and history offshore Norway

Records of submarine glacial landforms have been crucial for reconstructing the styles, extents, and dynamics of past glaciers and ice sheets across the many glaciated continental margins on Earth (Dowdeswell et al., 2016). A large number of landforms, from local to regional scale, have been observed and described offshore Norway, both on the seafloor and at depth. Many of these features are extensively covered in a recent compilation by Dowdeswell et al. (2016), so below we provide a brief summary of their geomorphological character and their use in ice sheet reconstruction along the Norwegian margin. In the following section the landform records are discussed in more detail.

5.1. Glacial landforms

5.1.1. Ice sheet flow indicators

A large number of streamlined landforms have been observed offshore Norway and can be used individually, or as part of a landform system, to determine ice flow direction, the evolution of ice sheet dynamics, and sediment transport pathways through a glacial cycle (Ottesen et al., 2005a; Dowdeswell et al., 2006). The different large-scale landforms include:

(i) Mega-scale glacial lineations (MSGL) are elongated corrugations 100–200 m wide, tens of metres high, and several kilometres long (Clark, 1993; Ely et al., 2016) (Fig. 13a). Due to their length and parallel concordance they are associated with fast-flowing ice (Clark, 1993; Stokes and Clark, 1999) and have been used to identify the locations of ice streams that operated along the Norwegian margin (Ottesen et al., 2005a, 2008; Andreassen and

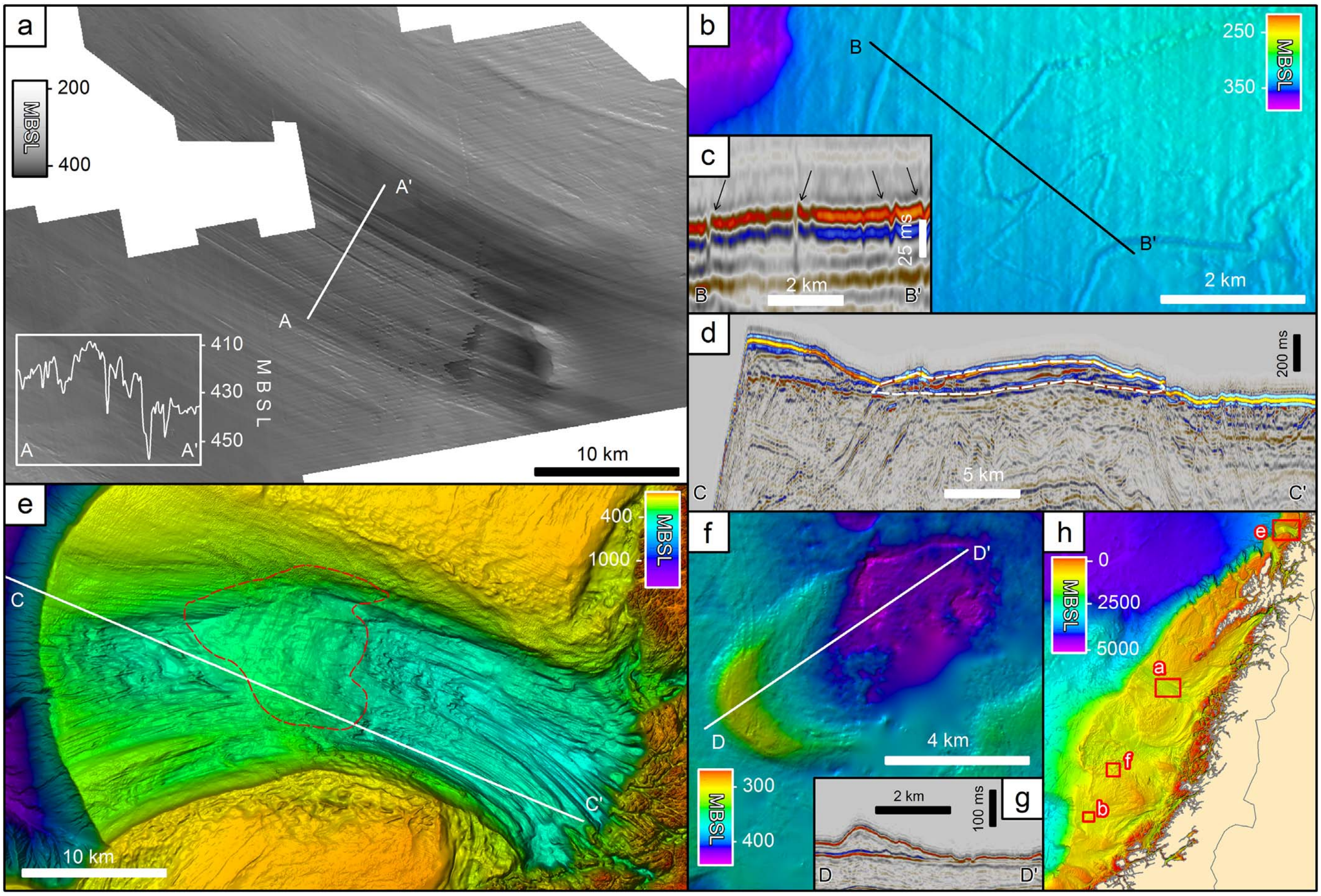

Fig. 13. Examples of several glacial landforms found offshore Norway. a) 3D seismic data from the seafloor of the Trænadjupet Trough show the presence of elongate landforms interpreted as MSGL. b) Example of iceberg scours on the seafloor from 3D seismic data. c) Seismic cross-section of the iceberg scours with arrows indicating berms of excavated material from the iceberg scour. Seismic line location shown in panel (b). d) Seismic cross-section with white dashed line showing a mid-shelf grounding-zone wedge. Location on panel (e). e) Seafloor bathymetry from the Malangsdjupet Trough with a prominent grounding-zone wedge highlighted with the red dashed line. Location of the seismic line in panel (d) is shown. f) Example of a hill hole pair from the Skjoldryggen area on the mid-Norwegian shelf. Seismic profile in panel (g) is indicated. g) Seismic profile showing the mound of excavated material from the glaciotectonic thrusting of an over-stiffened substrate. h) Location map of each of the landforms described. The bathymetric data shown in panel (e) is provided from the MAREANO multibeam data set (available at www.mareano.no) collected by the Norwegian Hydrographic Service.

Winsborrow, 2009).

(ii) Drumlins are smooth, streamlined oval-shaped hills that are typically 1 km long and up to 50 m high (Clark et al., 2009). Drumlin orientations indicate ice flow and have been used onshore and offshore Norway to reconstruct ice flow for the last glacial cycle (Ottesen et al., 2008; Winsborrow et al., 2010).

(iii) Crag-and-tail structures are small hills formed from resistant to erosion rocks that obstruct ice flow. This produces a tail of streamlined sediment on the lee side with typical elongation ratios of 3:1 to > 10:1. These features are observed on the northern Norwegian margin in areas where crystalline bedrock outcrops on the seafloor (Ottesen et al., 2008).

5.1.2. Ice sheet margin indicators

Landforms that provide direct evidence for the location of former ice sheet margins are widely used in the reconstruction of palaeoclimate and in particular the pattern of retreat of an ice sheet margin. Offshore Norway grounding-zone wedges (GZWs) and ice-marginal moraines are observed:

i) GZWs are formed as sediment accumulates over decades or centuries at the grounding-zone of a marine-terminating glacier during a stillstand (Dowdeswell et al., 2008; Batchelor and Dowdeswell, 2015). Wedges can reach several kilometres in length, but height is limited by the accommodation between the ice shelf and the preceding deposits (Batchelor and Dowdeswell, 2015) (Fig. 13d and e). GZWs have been identified along the entire Norwegian margin and indicate periodic ice margin stability during deglaciation after the LGM (Ottesen et al., 2008; Rydningen et al., 2013; Sejrup et al., 2016).

ii) Moraines, formed from an accumulation of unconsolidated debris at the ice sheet margin or proglacial glaciotectonic deformation, come in many different types (Barr and Lovell, 2014). End moraines, both terminal and recessional, form during ice margin stillstand and provide evidence for the pattern and direction of palaeo-ice flow. Terminal and recessional moraines have been used on several parts of the Norwegian margin to reconstruct the pattern of ice retreat (Ottesen et al., 2008; Winsborrow et al., 2010).

5.1.3. Ice sheet dynamics and hydrology

A large number of smaller-scale landforms have been identified along different parts of the Norwegian margin, each of which provides a snapshot into ice dynamics and subglacial hydrology at the time the features were formed. For example, iceberg scours are one of the most common features on a glaciated margin (Woodworth-Lynas et al., 1985; Dowdeswell and Bamber, 2007) (Fig. 13b and c) and have been used to reconstruct ancient ocean currents along the Norwegian margin (Newton et al., 2016). Glaciotectonic landforms such as hill-hole pairs (Ottesen et al., 2005a) have been observed along the Norwegian margin (Fig. 13f and g), as well as different types of folds, faults, and dislocations that can be seen in outcrops and on seismic reflection data (Huuse and Lykke-Andersen, 2000a; Pedersen, 2012). These data provide information on subglacial conditions, ice flow directions, and changes in ice stream dynamics during ice sheet growth and decay (Sættem et al., 1996; Huuse and Lykke-Andersen, 2000a; Winsborrow et al., 2016).

Landforms produced by meltwater processes have been observed along different parts of the Norwegian margin. Tunnel valleys are large, overdeepened valleys formed by subglacial meltwater (Ó Cofaigh, 1996). The valleys often start and stop abruptly with few or no tributaries and may be filled with glaciofluvial, glaciomarine, glaciolacustrine sediments, post-glacial sediments, or not infilled at all (Kristensen et al., 2007; van der Vegt et al., 2012; Moreau and Huuse, 2014). Tunnel valleys have been observed in the North and Barents Seas (Huuse and Lykke-Andersen, 2000b; Stewart and Lonergan, 2011; Bjarnadóttir et al., 2017; Newton and Huuse, 2017) but they are uncommon along the mid-Norwegian margin where the open-ocean setting may have prevented differential pressure build up. Eskers are formed from both subglacial and englacial ice-walled meltwater conduits in and under the glacier and are largely parallel to ice flow and are particularly common where ice rested directly on consolidated bedrock (Banerjee and McDonald, 1975). During ice margin retreat or downwasting, these channels leave behind long winding ridges of stream deposits. Observations of eskers are uncommon offshore Norway due to the soft sedimentary substrate and open-ocean setting.

5.2. LGM landform records and deglacial chronology

Recent work synthesising the large amount of literature on the deglaciation of the Fennoscandian Ice Sheet (FIS) has provided a step-change in our understanding of how ice sheet margins evolved through the deglaciation after the LGM. For readers interested in the finer details of this work we direct them to Hughes et al. (2016) and Stroeven et al. (2016) who reconstruct the FIS through the last glacial cycle. These authors used geomorphological data from features such as meltwater channels, striations and lineations, glaciofluvial deposits, and ice-marginal features combined with geochronological data derived from cosmogenic radionuclide, radiocarbon, and optically-stimulated luminescence dating to provide a detailed record of ice sheet history. Here, we provide a summary of the overall retreat patterns inferred from geomorphological and geochronological records of ice that was grounded along the Norwegian margin (Fig. 14).

During MIS 3 the ice cover which developed over Norway during MIS 4 had probably lost the majority of its volume and was likely restricted to only the highest altitudes (Wohlfarth, 2010). Following the Laschamp palaeomagnetic excursion (~41 ka), the ice cover, which is thought to have extended beyond the coastline of mid-Norway at this time, retreated during the Ålesund Insterstadial from 38 to 34 ka (Valen et al., 1995; Mangerud et al., 2010). Although the exact extent of the retreat is uncertain, the west coast of Norway is thought to have been ice free and, as a minimum, that mountain glaciation prevailed on the highest Scandinavian peaks (Hughes et al., 2016). After the interstadial, ice advanced across the west Norwegian coastline at the coastal town of Ålesund (Mangerud et al., 2010). This is correlated with evidence for growth of the Klintholm Ice Stream in eastern Denmark (Houmark-Nielsen, 2010), suggesting regional growth of the ice sheet (Hughes et al., 2016). Elsewhere, the extent of the FIS at this time is poorly constrained with limited evidence for ice cover in northern Sweden and Finland (Lagerbäck and Robertsson, 1988) and ice-free conditions in Finnmark (Olsen et al., 1996).

Concomitant with the global drop in relative sea level, the FIS cover was mainly terrestrially-based with the exception of ice that extended onto the continental shelf along the majority of the Norwegian margin (Ottesen et al., 2005a). After ~30 ka the margins of the European ice sheets varied dramatically as they coalesced or uncoupled, reached their maximum extents at different times in different areas, and retreated at different rates (Svendsen et al., 2004; Clark et al., 2012; Hughes et al., 2016; Stroeven et al., 2016).

The western margin of the FIS is thought to have remained close to the Norwegian coastline near Ålesund (Mangerud et al., 2010) at ~28 ka and to have retreated across Denmark and southern Sweden at the same time (Houmark-Nielsen and Kjær, 2003; Larsen et al., 2009). After this retreat, the FIS is thought to have advanced and become confluent with the British and Irish Ice Sheet over the North Sea at ~27–26 ka (Sejrup et al., 2009, 2016; Clark et al., 2012). On the mid-Norwegian shelf the ice likely extended to the shelf break sometime after 27.8–27.6 ka (Dahlgren and Vorren, 2003; Rørvik et al., 2010; Johnsen et al., 2012) (Fig. 14a).

The confluence of the Barents-Kara ice cover with ice over northern Norway, resulting in all three major European Ice Sheets being connected (Fig. 15a), occurred no earlier than ~25 ka (Hughes et al., 2016). By 25 ka, the glaciomarine sedimentary record and seafloor geomorphology suggest that the ice masses over the North Sea had already begun to separate in the northern North Sea (Sejrup et al., 1994, 2016). On the mid-Norwegian shelf, landform records and seismic stratigraphy show that the ice margin experienced several phases of advance and retreat to and from the shelf edge by up to 40 km (Dahlgren and Vorren, 2003; Brendryen et al., 2015). On the northern part of the margin, ice fluctuations near Andøya and along the mid-Norwegian shelf continued for ~2 kyr (Vorren et al., 1988; Nesje et al., 2007; Rørvik et al., 2010) before northern Andøya was deglaciated by ~22 ka (Alm, 1993; Vorren et al., 2015). IRD records suggest the Barents-Kara ice cover was still confluent with the ice in northern Norway at this time (Landvik et al., 1998).

From 22 to 18 ka major changes occurred along the Norwegian margin (Fig. 15b). A temporary re-advance occurred in the northern North Sea (Rise and Rokoengen, 1984; Sejrup et al., 2016) before growth and increased velocity of the Norwegian Channel Ice Stream led to drawdown of the ice divide (Nygård et al., 2007). This eventually led to margin retreat from ~20–19 ka (Svendsen et al., 2015) and separation of the two ice sheets sometime between 18.5 and 17 ka (Sejrup et al., 2016). On the mid-Norwegian shelf the timing of retreat is poorly constrained. Luminescence dating has been interpreted to suggest a substantial retreat of the ice margin at ~22 ka (Johnsen et al., 2012; Kolstrup and Olsen, 2012), but conflicting marine geophysical and geological evidence suggest that ice was still grounded on the shelf at a similar time (Elliot et al., 2001; Dahlgren and Vorren, 2003) (Fig. 14a). However, at this time the southern and eastern margins of the Fennoscandian Ice Sheet were close to their maximum extent (Larsen et al., 2006; Lasberg and Kalm, 2013) and it thus appears unlikely such a large retreat occurred on the western margin (Hughes et al., 2016).

As the British and Irish Ice Sheet separated from the FIS, the Barents-Kara Ice Sheet remained confluent (Fig. 15c) with ice over northern Norway. Seafloor geomorphology in the Barents Sea has shown that these ice masses separated as the ice retreated through the Bjørnøyrenna trough (Andreassen et al., 2014; Bjarnadóttir et al., 2014) (Fig. 14a). The retreat pattern and style is poorly resolved due to limited observational data, but it is thought to have occurred between 17 and 15 ka (Hughes et al., 2016; Newton and Huuse, 2017).

As deglaciation continued, moraines mapped offshore northern Norway indicate multiple stillstands during an overall retreat as ice moved closer to the coastline at ~16 ka (Stokes et al., 2014). Offshore southern Norway ice had retreated onshore at this time (Nygård et al., 2004; Sejrup et al., 2009). By ~15 ka the majority of the FIS had retreated behind the coastline along the Norwegian margin (Goehring

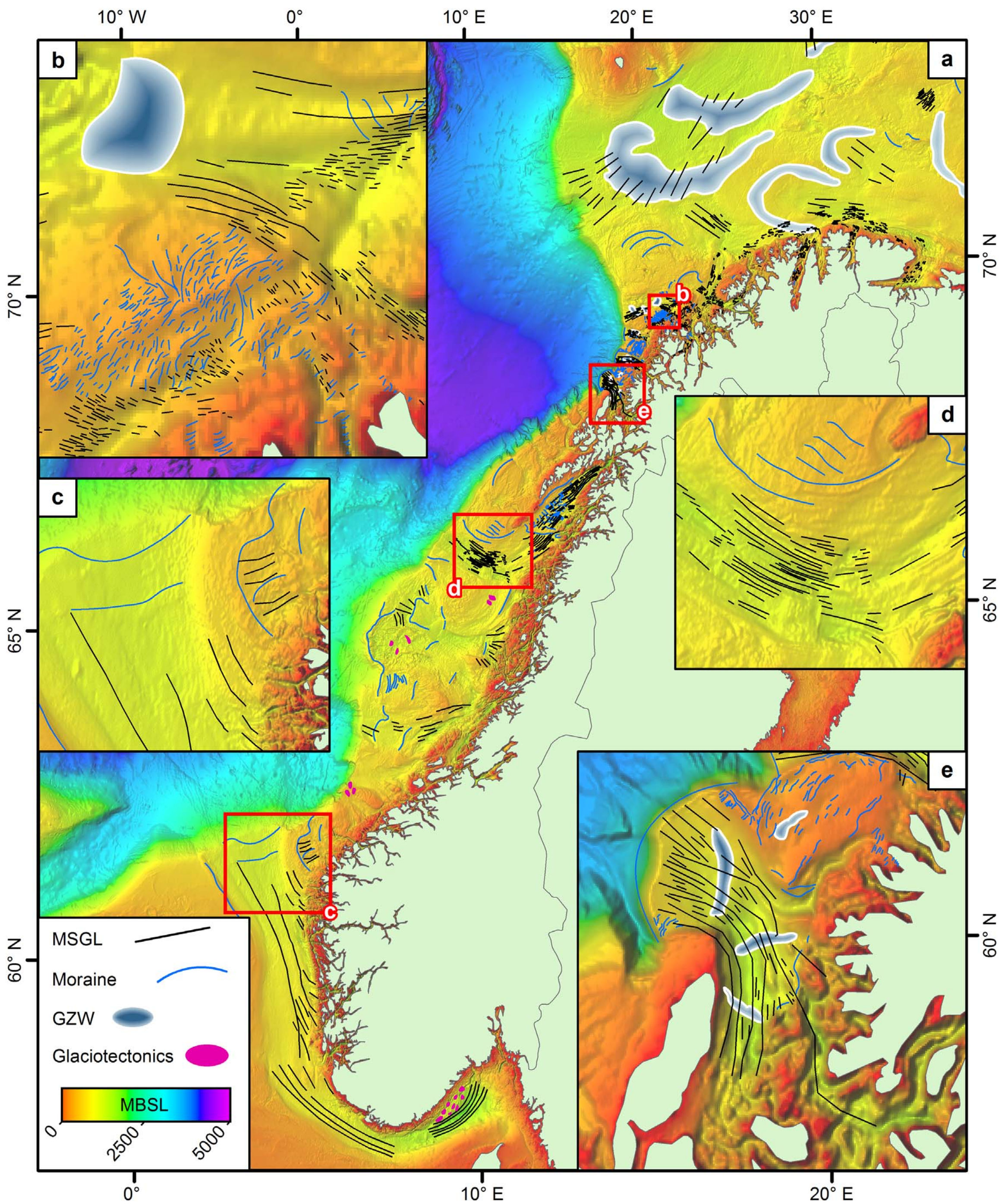

Fig. 14. Compilation of geomorphological data that have been collected along the Norwegian margin. a) An overview of the glacial landforms identified along the margin that have been used to interpret the evolution of the last Fennoscandian Ice Sheet. Locations of panels (b)–(e) are shown. b) Glacial lineations and moraine deposits on the northernmost part of the margin. c) Evidence of ice streaming and ice-marginal deposits at the mouth of the Norwegian Channel on the southern part of the Norwegian Atlantic margin. d) Landform record showing lateral moraines and glacial lineations in Vestfjorden on the northern part of the mid-Norwegian margin. e) Record of ice streaming and intermittent retreat of the ice cover in Andfjorden off of Troms. Figure compiled from glacial geomorphological studies along the entire margin (Elverhøi et al., 1993; Ottesen et al., 2005a, 2005b, 2008, 2016; Laberg et al., 2007, 2009; Andreassen et al., 2008, 2014; Andreassen and Winsborrow, 2009; Winsborrow et al., 2010, 2011; Rüther et al., 2011; Bjarnadóttir et al., 2013, 2014; Rydningen et al., 2013; Newton and Huuse, 2017). Note the very large grounding zone wedges in the Barents Sea and the MSGL that are present in several of the cross-shelf troughs on the mid-Norwegian shelf. The area north of Troms has seen significantly more study due to the MAREANO programme which has collected a large amount of high resolution multibeam data.

et al., 2008; Hughes et al., 2016), with the exception of northern Norway. Glaciomarine sedimentation in the Barents Sea at ~15–14.4 ka and lake records indicate deglaciation occurred later, at ~14 ka (Romundset et al., 2011; Rüther et al., 2011) despite evidence suggesting thinning of ice over northern Norway (Fjellanger et al., 2006). By 14–13 ka the FIS had retreated onshore along the entire Atlantic coastline of Norway (Johnsen et al., 2009; Romundset et al., 2011).

Prior to the Younger Dryas (c. 12.9 ka), the ice margins have been reconstructed well inland of the moraines associated with the ~12 ka

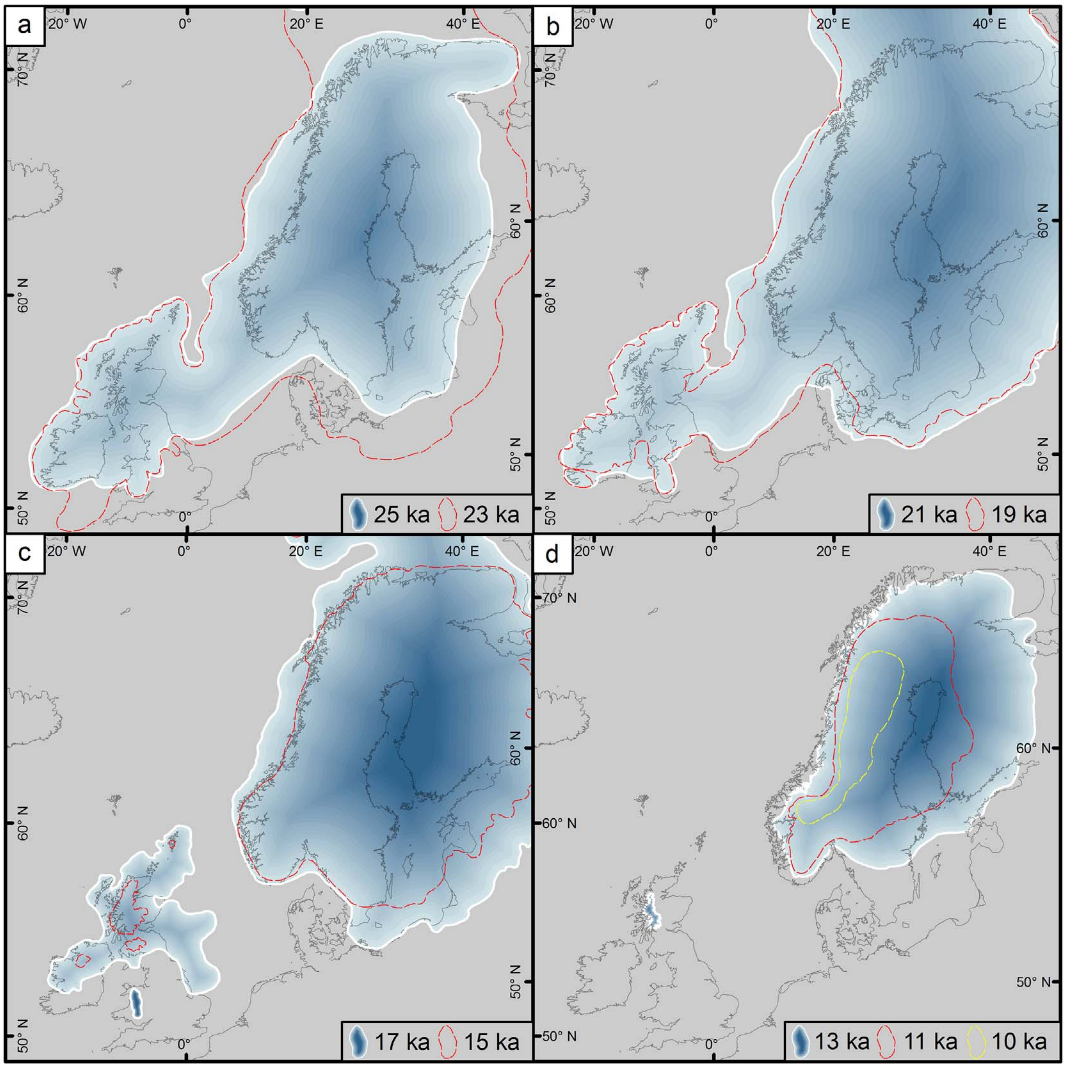

Fig. 15. a–d) A selection of time slices of the British and Irish Ice Sheet, the Fennoscandian Ice Sheet, and the Barents-Kara Sea Ice Sheet through the deglaciation after the last glacial maximum. Note how the ice sheet is at its maximum extent at different times in different sectors. Data compiled from the 'most-credible reconstructions' from DATED-1 (Hughes et al., 2016).

re-advance (Mangerud et al., 2011; Hughes et al., 2016; Stroeven et al., 2016) (Fig. 15d). After the Younger Dryas the ice margin was losing mass both laterally and vertically resulting in a heavily dissected ice cover split into numerous local ice covers (Linge et al., 2006; Linge et al., 2007; Goehring et al., 2008). The final deglaciation of the Fennoscandian Ice Sheet is thought to have occurred at ~10–9 ka when ice rapidly melted away in the deep valleys located near to the former ice divide (Hughes et al., 2016; Stroeven et al., 2016).

The record of deglaciation described above and displayed in Fig. 15, has shown that the pattern of deglaciation varied across Fennoscandia as ice retreat operated asynchronously. This likely related to factors such as: mass balance and distribution of precipitation (e.g. Clason et al., 2016), variability in substrate conditions (e.g. Winsborrow et al., 2016), and the coeval effects that these patterns had on ice sheet thermal conditions and ice dynamics. Large parts of the FIS were grounded below sea level when at its maximum extent and this interaction between terrestrial- and marine-terminating ice would have had a fundamental control on ice sheet evolution.

This combination of factors likely led to the variable responses of the European ice cover through the deglaciation. For example, as deglaciation progressed, penetration of warmer North Atlantic waters and atmospheric temperatures could have helped to drive melt above and below the marine-terminating part of the FIS. During Heinrich events, regional cooling could have temporarily subdued mass loss, resulting in changeable rates of melt and retreat. Likewise, the geological setting and morphology of the shelf would have influenced the rate of retreat. Along the southern and northern parts of the margin, the LGM ice sheet would have had a shorter distance to travel from its maximum extent at the shelf edge to the coast. Once ice retreated behind the coastline the direct influence of ocean warming would have been reduced as atmospheric warming became the dominant control on mass loss.

5.3. Pre-LGM landform records

Along the Norwegian margin considerable work has been carried out investigating landform records to reconstruct the extent and dynamics of the FIS during the last glacial cycle. By comparison, very little is known about whether older landform records are present within the Plio-Pleistocene stratigraphy and what they can tell us about older glaciations. Most of the work looking at older records has used seismic stratigraphy and core analysis to infer the presence of grounded or floating ice along the Norwegian margin (Dahlgren et al., 2002;

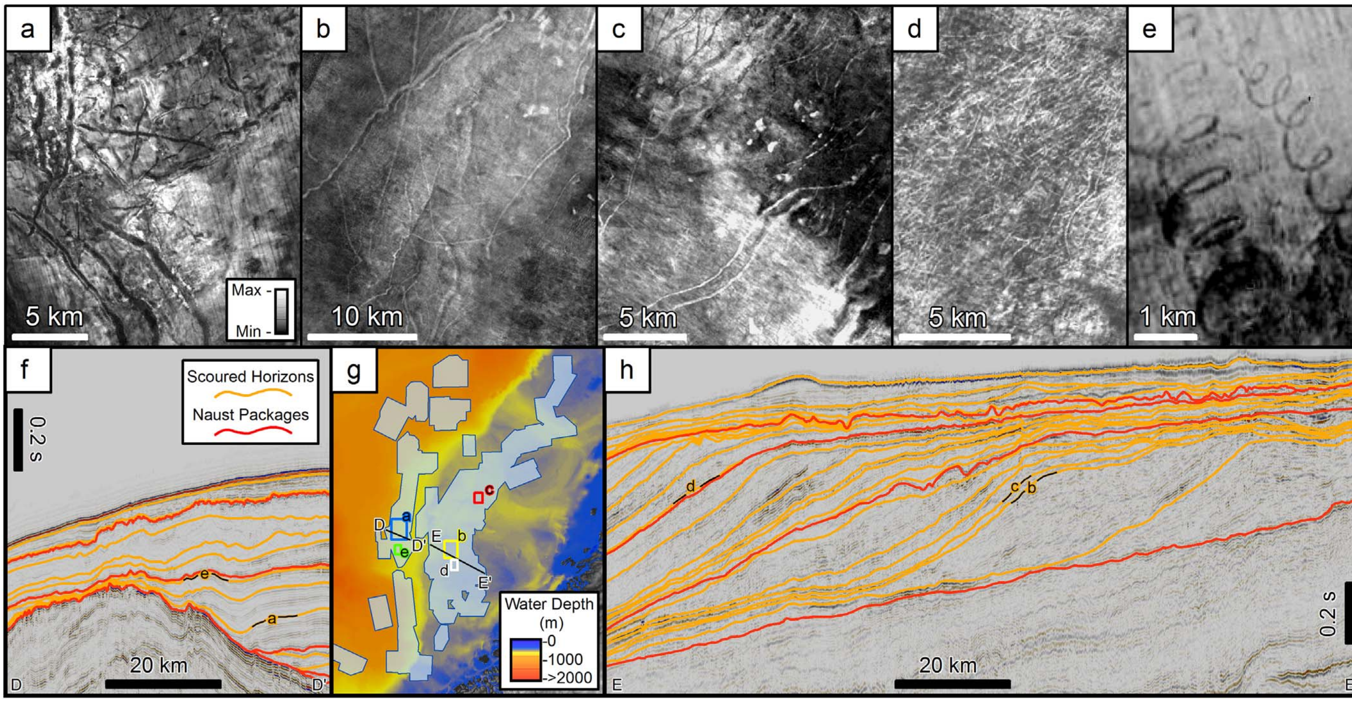

Fig. 16. Examples of buried iceberg scoured palaeo-seafloors. Frequent episodes of iceberg scouring suggests that marine-terminating ice, whether it was floating or grounded, was common throughout the Pleistocene along the Norwegian margin. Panels (a) to (e) show examples of iceberg scouring, whilst panels (f) to (h) show the geographic and stratigraphic location of the palaeo-seafloors.

Hjelstuen et al., 2005) without being able to show direct evidence of landform assemblages. Improvements in seismic image quality and coverage, and in particular the availability of industry 3D seismic data have provided a step-change increase in our ability to image buried glaciogenic landforms in their stratigraphic context.

Until recently only a limited amount of work had investigated palaeo-seafloors for evidence of glacial landforms, with iceberg scours and MSGL being the most common. Iceberg scours have been observed at the top of the Naust Formation S-unit (Hjelstuen et al., 2004b) and from within the N-unit (Ottesen et al., 2009). Iceberg scours identified within the upper part of unit-N are about 350 m below the modern seafloor and are dated to ~2 Ma during one of the earliest Pleistocene glaciations (Ottesen et al., 2009). More recent work has shown that an extensive record of iceberg scouring exists on the mid-Norwegian shelf where iceberg scours have been observed throughout the five main Naust Formation units (Fig. 16) (Huuse and Newton, 2015; Montelli et al., 2017). If the oceanography at the LGM (Death et al., 2006) is a reasonable approximation of ocean current patterns throughout the Pleistocene glaciations, then it is likely that the icebergs were sourced locally and had not travelled across the North Atlantic from the Greenland or North American Ice Sheets. This would suggest that ice cover along the Norwegian margin repeatedly grew large enough to extend beyond the fjords and onto the continental shelf through most glacials of the Pleistocene.

Icebergs would also have been calved from glaciers contained within fjords, but these icebergs would have been limited in their draft by the fjords being significantly shallower than the present, and sea level regressions being reduced in amplitude compared to the Late Pleistocene. Thus, icebergs sourced from the fjords are unlikely to have been sufficiently deep enough to repeatedly scour the continental slope off of mid-Norway through the Early Pleistocene, thus meaning that ice sheets likely extended beyond the fjords and the Norwegian coastline (Huuse and Newton, 2015; Montelli et al., 2017).

On the mid-Norwegian margin 3D seismic data have also shown the presence of several sets of MSGL through the last three major glaciations (Dowdeswell et al., 2006; Montelli et al., 2017). The MSGL record and patterns of sediment deposition show that the ice sheets underwent major changes in ice flow directions in response to changes in the topography of the continental shelf through the different glacial periods (Dowdeswell et al., 2006).

Evidence of grounded ice within the older glacial stratigraphy of the Norwegian margin is scarce. The only previous work showed evidence for some of the oldest records of buried MSGL on the mid-Norwegian margin (Ottesen et al., 2009; Montelli et al., 2017). MSGL were observed on the top of the N-unit in southern Haltenbanken and have been dated to ~1.5 Ma (Ottesen et al., 2009) (Fig. 17a). A second set of MSGL were observed within the Vestfjorden area and have been tentatively dated to ~1.6–1.9 Ma (Montelli et al., 2017). Within the A- to T-units on the mid-Norwegian margin and also within the Mid to Late Pleistocene stratigraphy of the Norwegian Channel (Rise et al., 2004) and offshore Troms (Rydningen et al., 2016), MSGL observations become increasingly common, particularly after ~800 ka when the length of the glacials extends to 100 kyr cycles (Dowdeswell et al., 2006; Huuse and Newton, 2015; Montelli et al., 2017) (Fig. 17b and c). There are limited observations of GZWs within the S- and T-units and retreat moraines along the margin that have been observed within the three youngest units that provide evidence for grounded ice in the Middle and Late Pleistocene (Montelli et al., 2017). The majority of evidence supporting the presence of grounded ice on the Norwegian margin is located in the central part of the margin rather than to the north and south. This suggests that the subglacial record is better preserved here, or, more likely, that the narrow continental shelves in the north and south resulted in most sediment bypassing the shelf. Without burial of the landforms on the shelf, the preservation of features such as MSGL is limited compared to the middle part of the margin where a large number of palaeo-shelves have been preserved.

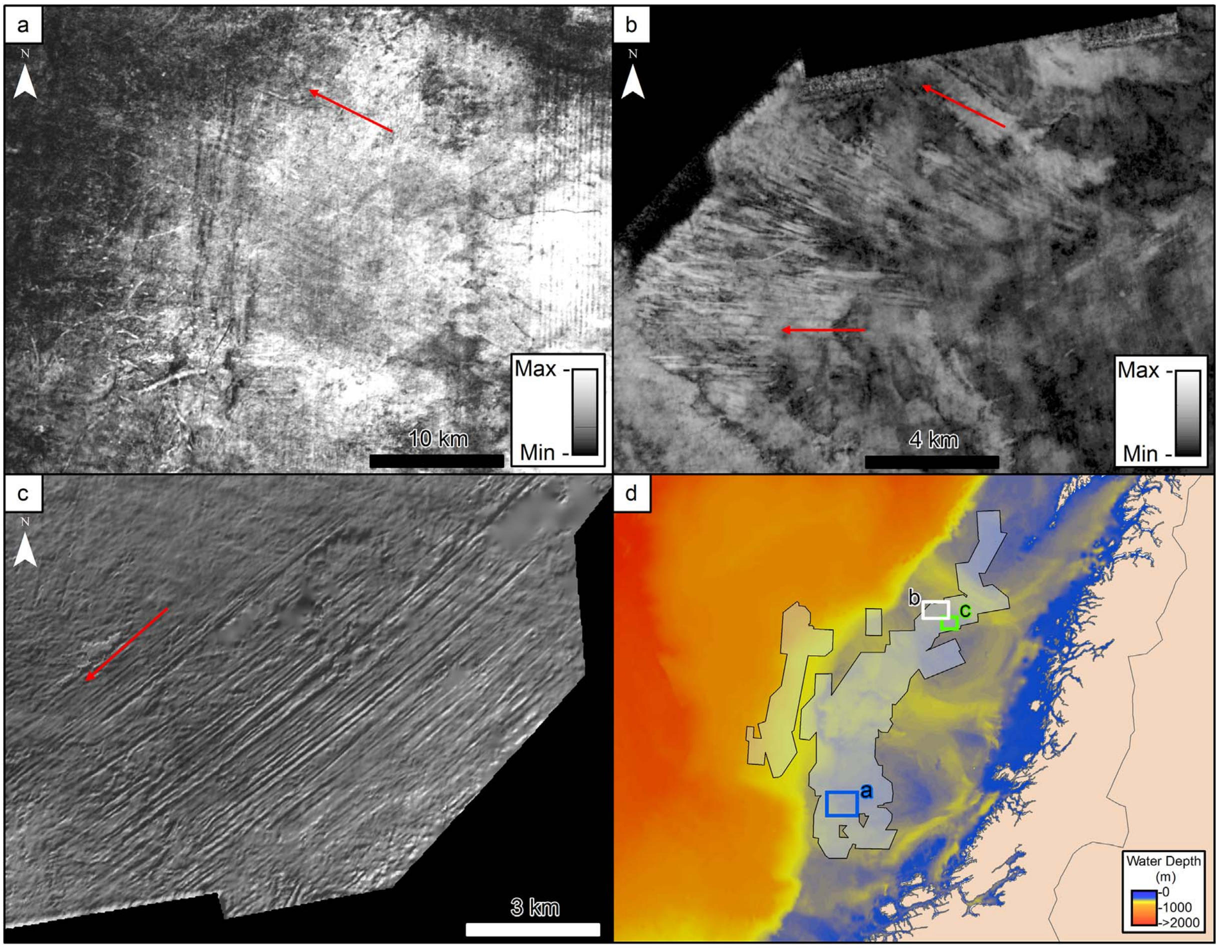

Fig. 17. Examples of buried mega-scale glacial lineations (MSGL) within the Naust Formation. a) MSGL in southern Haltenbanken at the top of the Naust N unit and have been dated to ~1.5 Ma by Ottesen et al. (2009). b) MSGL within the Naust A unit. c) Shaded relief image of MIS 6 MSGL on the northern part of the mid-Norwegian margin showing that ice flow was different compared to the LGM (Dowdeswell et al., 2006). d) The location of the features identified using the 3D seismic reflection data. In all panels the red arrows are parallel to the inferred direction of ice flow indicated by the MSGL.

6. Future research directions on the Norwegian margin

From the discussion above, it is clear that the Norwegian margin contains a long-term record of oceanic, climatic, and terrestrial environmental changes through the late Cenozoic. These records have been studied across different timescales because of the margin's key, mid- to high-latitude location and its position within the Northeast Atlantic part of the global ocean circulation. An improved understanding of changes along the Norwegian margin would lead to a better understanding of local and regional climatic changes in the Northeast Atlantic, including heat transport from low-latitudes to the Arctic, and ocean overturning in the North Atlantic. The extensive database of geophysical data along the margin has yielded consistent records on long-term environmental changes through the Pleistocene. Although the seismic stratigraphic framework is tied to local core data (e.g. ODP Sites 642–644), the geochronology of the older Naust Formation units remains very poorly constrained, which severely limits the analysis and interpretations of the changes that these sedimentary archives preserve. Thus, although the offshore records of the Norwegian margin have been well-studied, significant additional insight could yet be derived from better dating of the sedimentary archives. In this section we outline potential avenues of future research and possible areas of interest for calibration of the stratigraphy by ocean drilling.

Much of the research discussed in this review paper falls broadly under one scientific aim, that is, to develop a better understanding of how changes in boundary conditions, whether that be the entire Earth system or components within it, can be derived from changes in the sedimentary archives preserved offshore. Firstly, the nature of glaciation on a passive oceanic margin means that the cycling through different glacial-interglacial periods has removed and reworked some of the sediments deposited on or proximal to the continental shelf, resulting in heterogeneous preservation and complicating their interpretation. Secondly, the paucity of geochronological control means that the expanded and largely continuous stratigraphic record present in the glacial clinoforms and distal contourites is uncalibrated and any attempts to link it with global proxy records are associated with significant uncertainty.

Extensive seismic data along the Norwegian margin means that the locations of sediment reworking and removal can be identified (e.g. Fig. 7 from Montelli et al. (2017)). The architecture and morphology of reworked deposits and unconformities can also provide insight into the processes and environmental conditions that caused these changes. For example, an unconformity with glacial lineations provides insight into the mechanism of erosion and the sediment pathway. These records provide a snapshot into environmental changes across long time-scales and can be used to test the ability of long-term climate models to reproduce these observations across multiple glacial-interglacial cycles.

The extensive availability of seismic reflection data and borehole data from industry operations (including wireline logs and cuttings or side-wall core data) would allow for the pin-pointing of the best potential locations for future ocean drilling to calibrate this important archive. If successful, this could go a long way to resolving the lack of age control described above. The preserved palaeo-shelves in the upper part of the stratigraphy could be one potential target for drilling, whilst

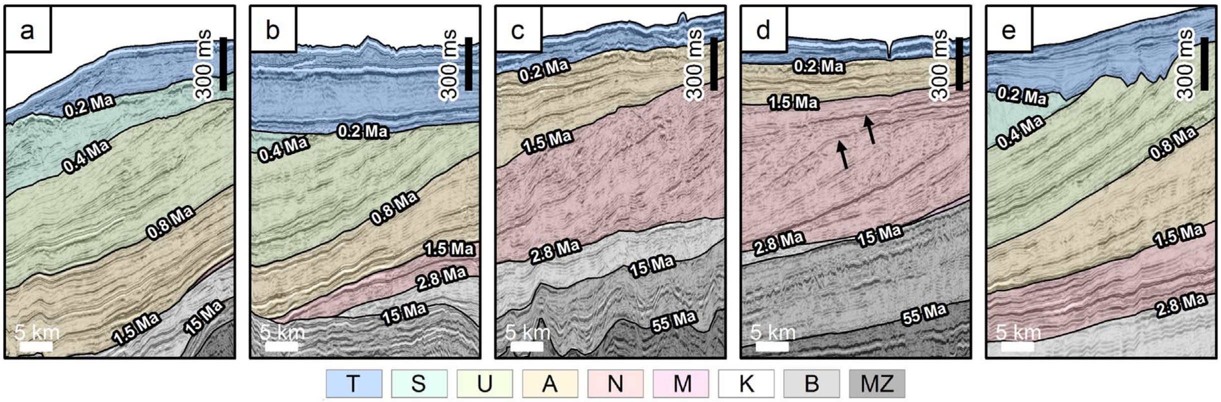

Fig. 18. a–e) Several possible areas of interest for future drilling along the Norwegian margin. Each panel can be located on Fig. 3. Drilling in areas such as these could yield records throughout the Cenozoic, whilst targeting sediments from slope deposits, unconformities, and buried palaeo-shelves (see arrows on panel (d)) that are known to contain glaciogenic landforms could provide crucial chronology to environmental changes in the latest Cenozoic. Stratigraphic abbreviations are the same as Fig. 1.

the ice-distal portions of the stratigraphy (e.g. the prograding clinoforms) (Fig. 18) would also provide an integrated record that goes well beyond the LGM and through the late Cenozoic. Although ice-proximal deposits are difficult to date, if integrated with dating derived from the more distal contourite deposits, this combination has the potential to provide time constraints to records of environmental changes documented using seismic data. Even if only a few dates of key events such as the Early Pleistocene evidence of grounded ice (Ottesen et al., 2009; Montelli et al., 2017) could be extracted, this would have important implications for our understanding of long-term climate changes along the Norwegian margin.

ODP Sites 642–644 have been extensively used to investigate palaeo-environmental changes in the Norwegian Sea (Fig. 7). However, new information could be extracted from these existing cores or from additional coring sites in the Norwegian Sea using new and evolving dating techniques (Balco and Rovey, 2010; Simon et al., 2012), sea surface temperature reconstructions (Kucera et al., 2005), and biomarker analysis (such as the sea ice proxy IP_{25}) (Belt et al., 2007). Each of these areas of research has the potential to provide significant insight and chronology across different long-term and short-term timescales.

A detailed seismic stratigraphic and geomorphological framework is already in place, and if it were to be constrained with geochronology, lithology changes, and rock properties, such records could be used to convert seismic volumes to rates of change. This would improve correlation with records from other margins and epicontinental seas. A high-resolution record would also provide a means for quantifying large-scale rates of erosion and deposition through different periods of glaciation. This would feed back into the analysis of long-term source-to-sink dynamics and the efficacy of glacial processes to induce topographic changes. This has the potential to provide a step-change increase in our understanding of environmental changes in the region and more generically for understanding the evolution of glaciated margins with mountainous hinterlands.

7. Conclusions

The Norwegian margin has seen dramatic environmental changes throughout the late Cenozoic and in particular during the Plio-Pleistocene. In this review we have brought together a large volume of literature that covers environmental changes in order to fully understand the causes, characteristics, and consequences of these changes in a holistic way. This review concentrated on the large-scale geological evolution and concomitant changes and feedbacks that have occurred in the hydro-, atmo-, and cryosphere across both long-term (~Myr) and short-term (~kyr) timescales. Erosion and deposition along the margin have been controlled by both ocean currents and ice sheets. Contourites along the Norwegian margin have provided insight into the connection of the Arctic and Atlantic Oceans from the Miocene onward. The ~1 km thick succession of the late Pliocene and Pleistocene Naust Formation has extended the shelf by up to 150 km and this can be linked to the grounding of ice sheets on the continental shelf. Seismic reflection data has led to observations of glacial landforms such as mega-scale glacial lineations on clinoform topsets in the Early Pleistocene. Glaciogenic debris flows have also been observed as a key depositional process at the shelf edge and have contributed to the generation of erosive turbidity currents and canyon formation. It is clear from this review that the Norwegian margin is a crucial region for understanding late Cenozoic environmental changes because it provides an important juncture for climate changes between lower latitudes and the Arctic.

The processes operating along the margin are relatively well-understood, but the lack of robust and detailed age control means that it is difficult to use these observations for constraining hindcast numerical models of climate evolution. With the benefit of an almost continuous coverage of seismic reflection data, additional geochronology would allow us to better understand the rates, styles, and feedbacks of environmental changes through the late Cenozoic. Due to its location, further ocean drilling along the Atlantic margin of Norway could yield information that is of global significance. Better constrained records of contourites, glacial landforms, and unconformities described in this review would allow for a more robust assessment of thresholds and forcings in the ocean-climate system of the Northern Hemisphere.

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