Glacial landforms and Quaternary landscape development in Norway

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

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Introduction

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

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

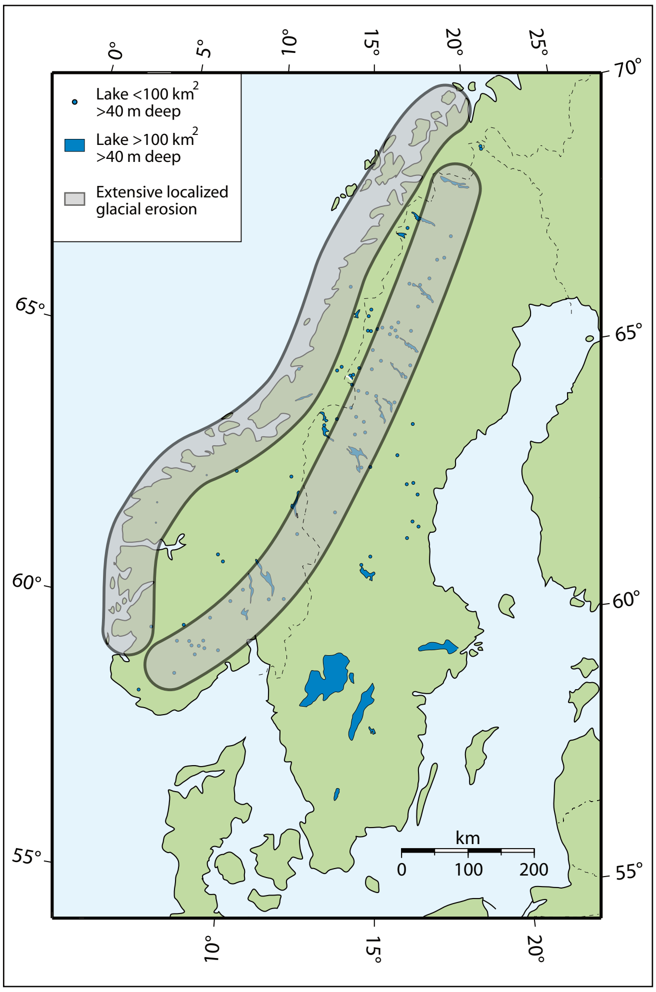


Figure 1. First-order glacial erosion landforms are centred around the Scandinavian mountain chain with fjords on the Norwegian coast and deep piedmont lakes east of the mountains

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

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

Pre-Quaternary landforms

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

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

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

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

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

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

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

Glacial landforms with some examples from Norway

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

Glacial erosional landforms

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

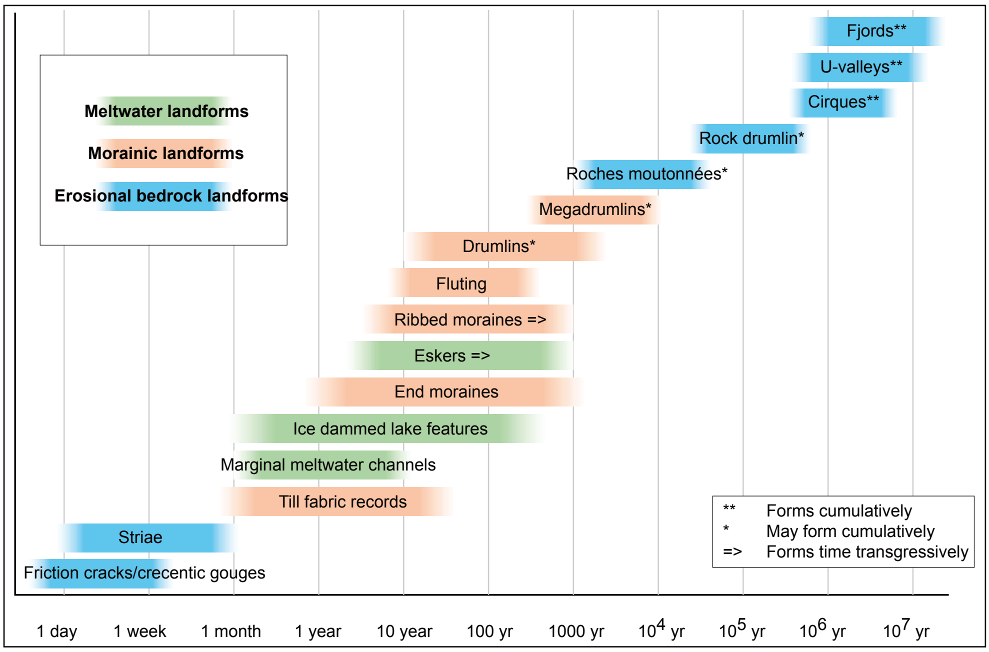


Figure 3. Glacial erosional and depositional landforms, their relative size and required time of formation. Copyright Clas Hättestrand, Stockholm University.

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

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

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

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

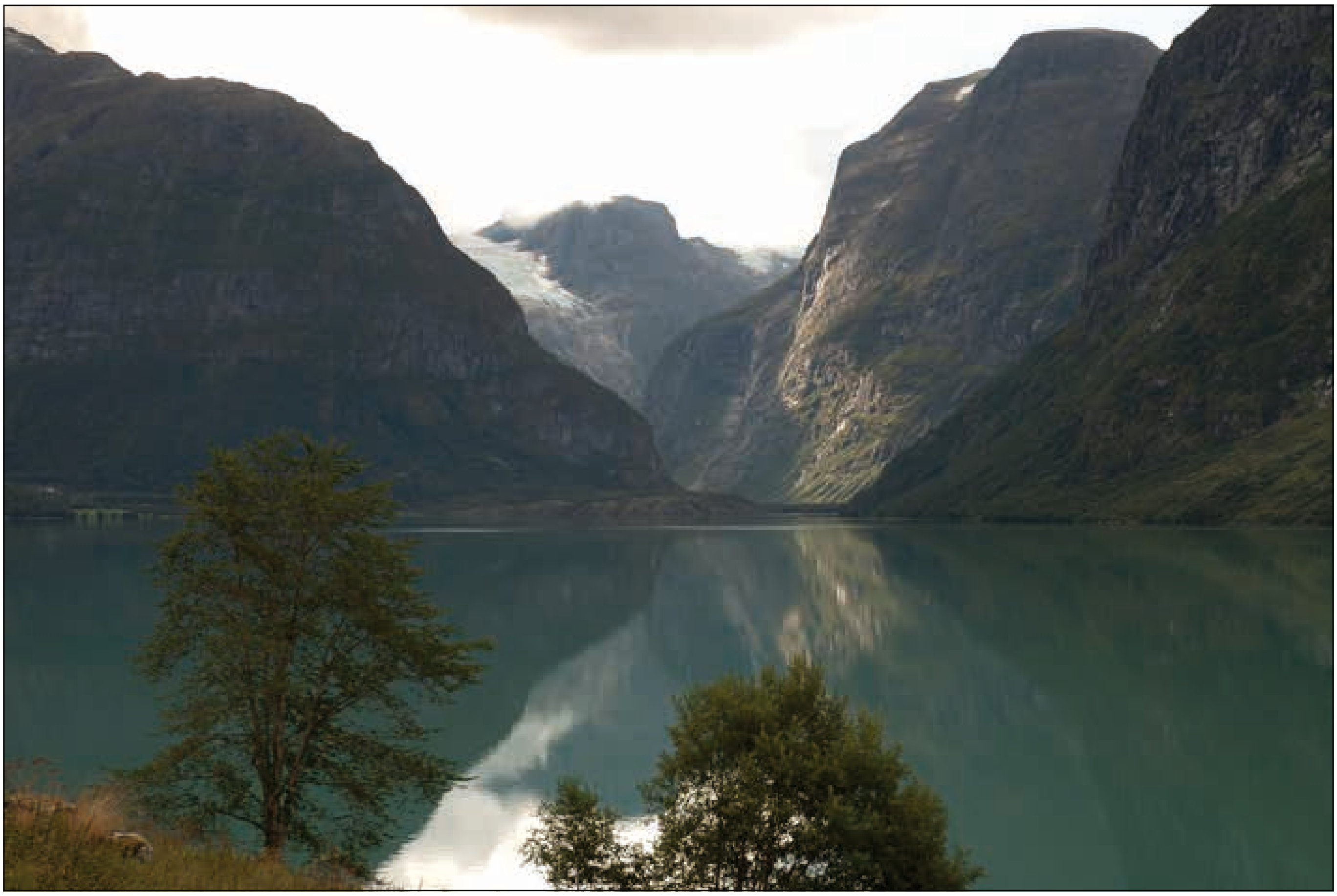


Figure 4. U-shaped valley with overdeepened lake basin at lake Loen close to Nordfjord, western Norway.

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

Glaciofluvial erosional landforms

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

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

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

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

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

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

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

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

Glacial depositional landforms with examples from Norway

Glacial depositional landforms

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

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

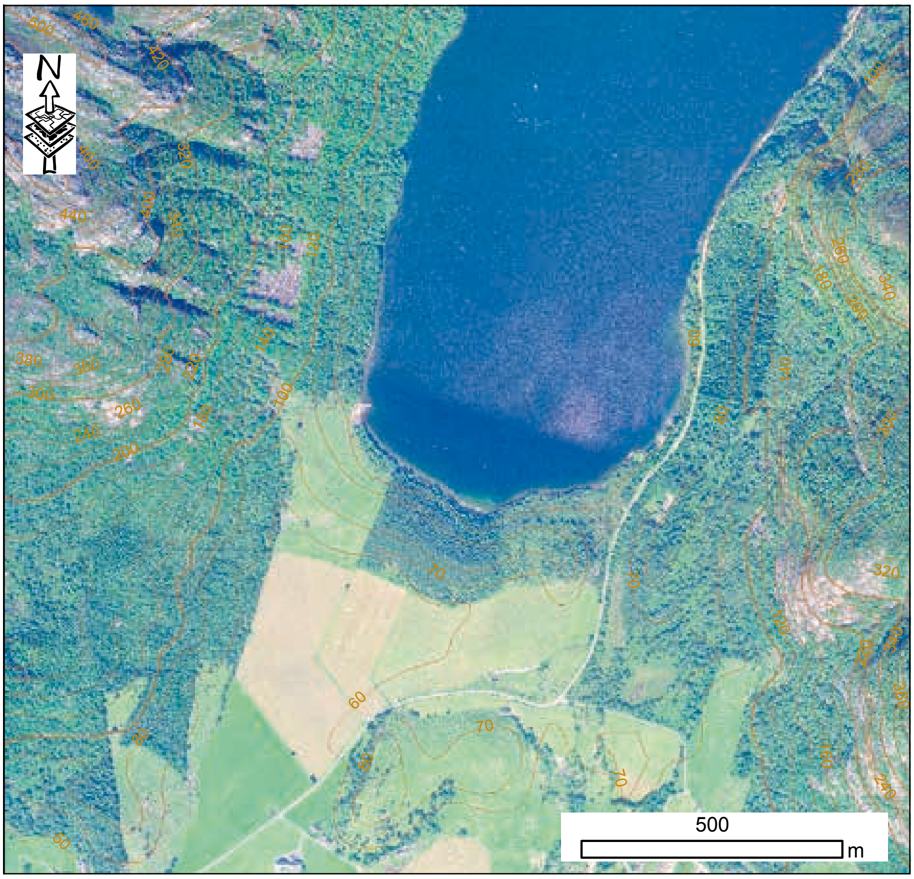


Figure 7. Aerial photograph of the Vassrygg Younger Dryas end moraine and proglacial outwash, situated just south of the Haukalivatnet.

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

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

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

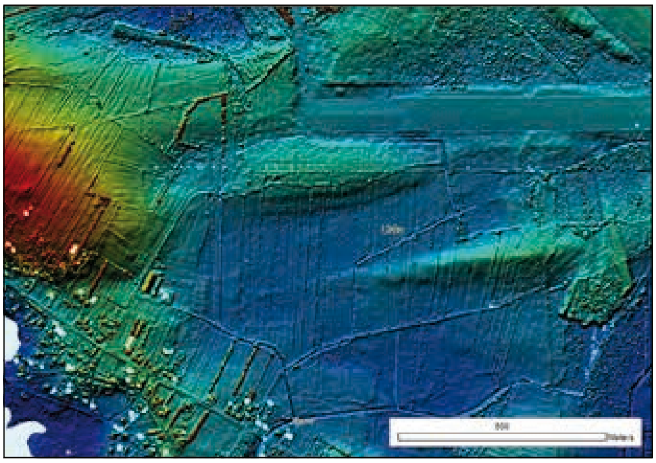


Figure 10. Drumlins on Lista in South Norway shown in a shaded relief map produced through high-resolution LiDAR digital-elevation data. The drumlins show ice flow from ENE towards the Skagerrak ice stream.

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

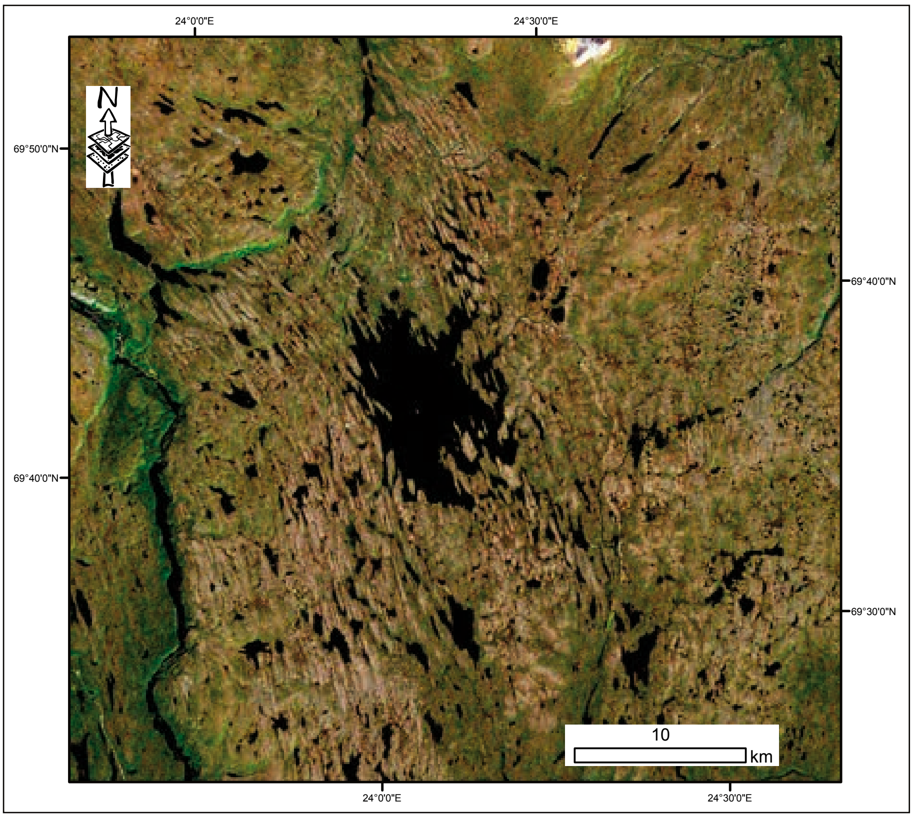


Figure 11. An extensive drumlin swarm on Finnmarksvidda, northern Norway, depicted in a Landsat 7 satellite scene. The drumlin swarm 'bends' due to migration of the ice-dispersal centre during the last deglaciation, thus showing the time-transgressive nature of drumlin formation. Also note the esker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

Glaciofluvial depositional landforms

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

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

Glacial landforms through the last glacial cycle

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

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

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

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

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

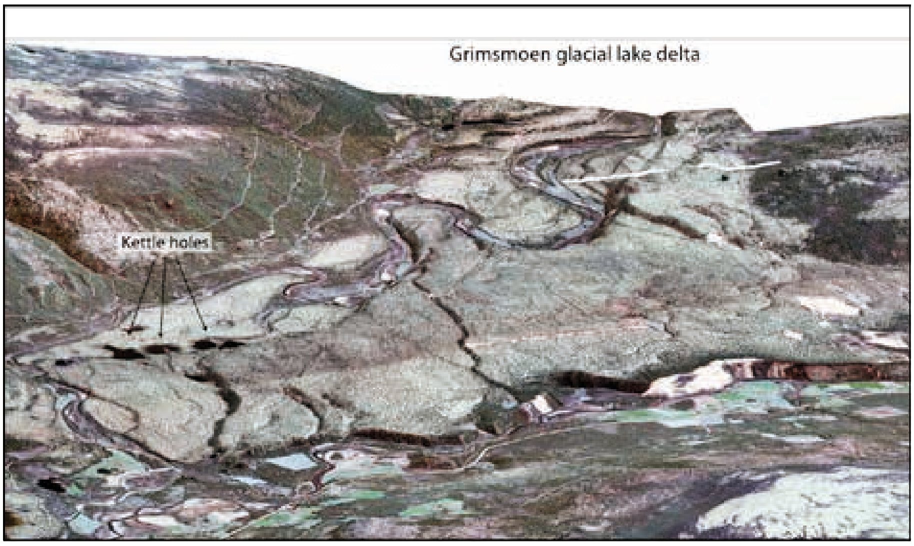


Figure 12. Aerial photographs draped over a digital elevation model (topography) showing a 3D view of the Grimsmoen glacial-lake delta in south-central Norway. Note the different terrace levels indicating lowering of the glacial lake during deposition. Kettle holes are evidence for stranded icebergs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

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

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

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

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

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

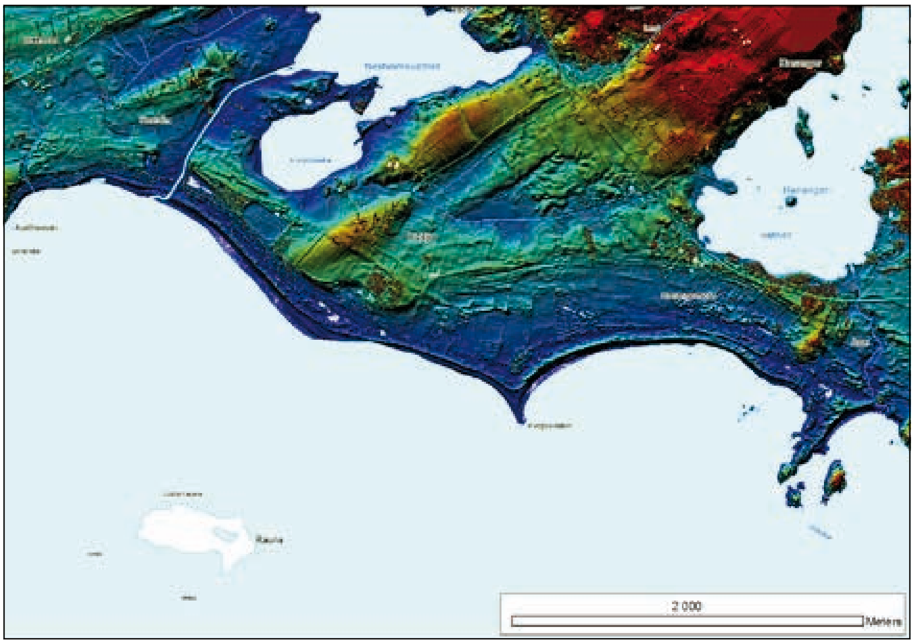


Figure 13. Shaded relief map showing large drumlins and multiple fossil shorelines on Lista, southernmost Norway. Note the sequence of shorelines between Kviljo and Hanangermona. The map is produced from high-resolution LiDAR digital elevation data.

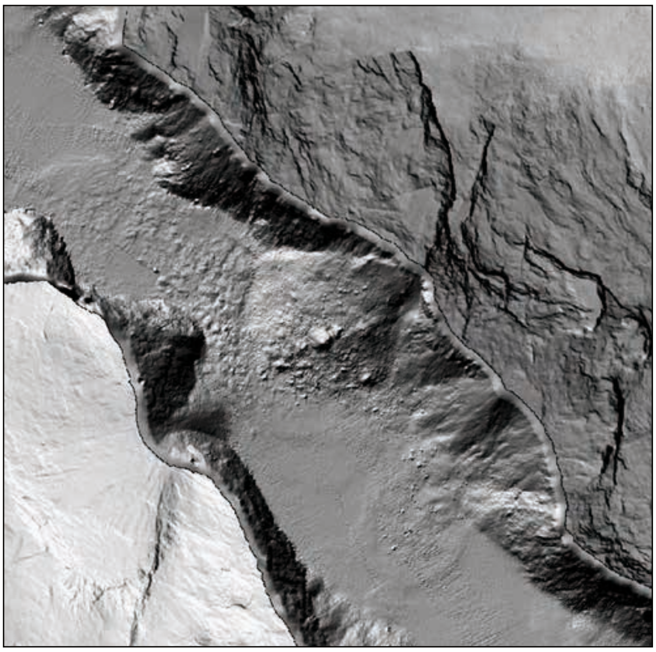


Figure 14. Shaded relief in a combined bathymetric and topographic dataset showing the floor of Tafjord. Note the very large submarine rock-avalanche deposits in the fjord.

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

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

Postglacial landforms— the Holocene landscape development

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

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

Marine landforms

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

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

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

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

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

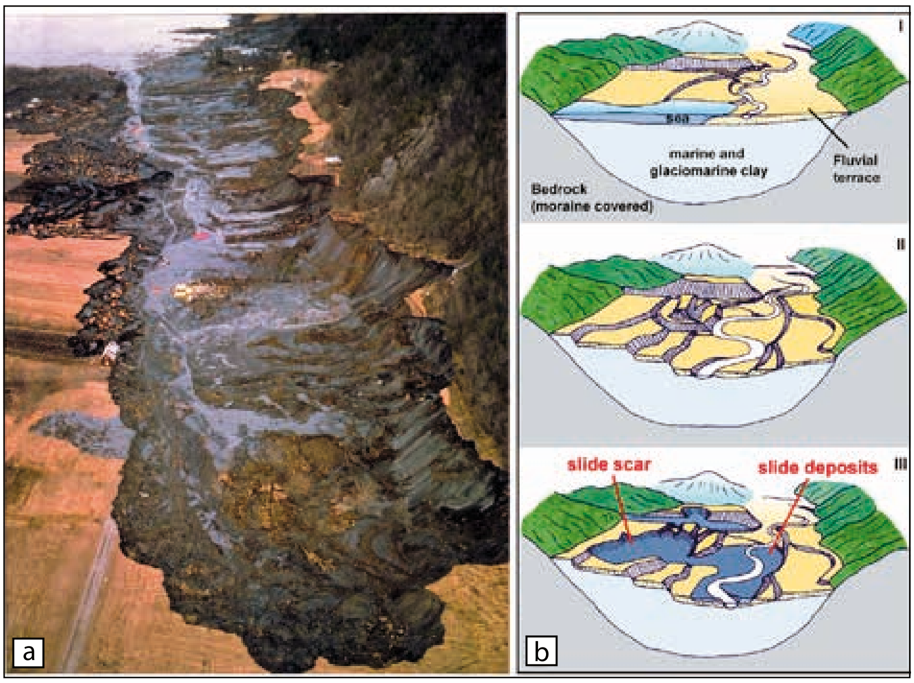


Figure 17. (a) Press photograph of a quick-clay slide at Rissa, mid-Norway. (b) Cartoon showing the evolution of quick-clay slides. Shortly after deglaciation isostatic rebound lifts the landscape causing fluvial incision, which in turn might trigger quick-clay slides.

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

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

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

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

Gravitational processes

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

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

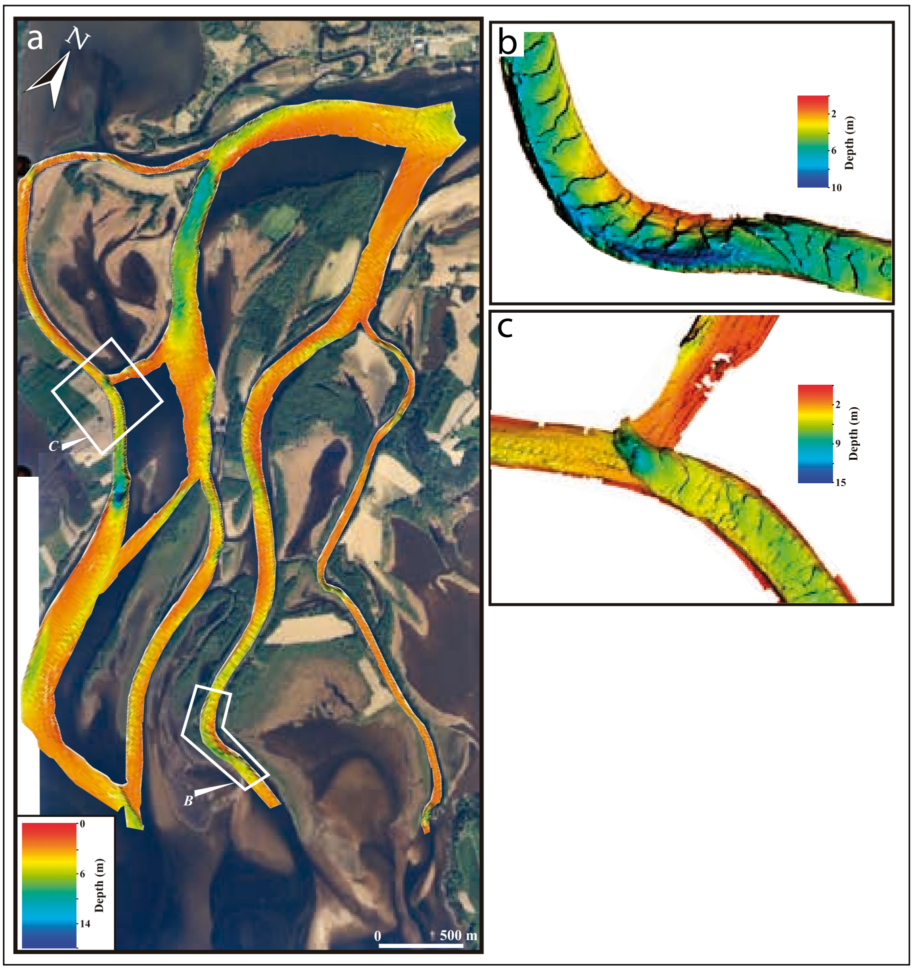
will cause increased overburden stresses in the rock mass. Second, deglaciation (glacier melting) of a landscape causes debuttressing when the supporting ice melts. Unloading of glacier ice may thus cause release of stress, which was built up in the rock mass during glacial loading (Augustinus 1992,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

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

Fluvial processes

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

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



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