

# **Late-Pleistocene and Holocene mountain permafrost geomorphology of Norway and Iceland**

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# Abstract

The combined effect of glacial and periglacial processes on landscape evolution has recently been termed the ‘cryoconditioning’ of landscapes, and largely affect the Fennoscandian landmass. Further, the distribution of permafrost both temporally and spatially during and after the last glaciation affect the overall geomorphic expression. In this thesis the product of landscape evolution is investigated in terms of the geomorphic imprint of glacial and periglacial processes, where the interactions between glaciers and permafrost have been particularly focussed upon. Whereas the glacial variations over the Holocene are relatively well known the same is not the case for the permafrost distribution.

As a starting point, inventories of landforms indicating present and former permafrost were compiled for mainland Norway and northern Iceland. The main findings from the inventories were (1) a low abundance of landforms in Norway and a high abundance of landforms in Iceland probably due to differences of bedrock competence, (2) an apparent change in processes leading to rock glacier formation occurred in mid-Holocene from a dry, periglacial regime characterizing early-Holocene to a humid, glacially-dominated regime in mid- to late-Holocene, and (3) warm and maritime permafrost regions are dominated by permafrost landforms formed by the influence of glaciers. For Iceland the occurrence of relict rock glaciers at sea level imply a possible earlier deglaciation or alternatively a less extensive Last Glacial Maximum (LGM) than commonly accepted.

As a second step, modelling of permafrost variations over the Holocene was performed in depth and spatially, using a 1D heat flow model and a 2D equilibrium model (CryoGRID1.0). During the warm Holocene Thermal Maximum (HTM) the permafrost survived at high altitudes in southern Norway, whereas during the ‘Little Ice Age’ (LIA) the permafrost reached its greatest extent both in depth and spatially. From these results altitudinal zones of permafrost ages was suggested, analogous to age patterns for Arctic

permafrost. From the Neoglaciation until present, the potential of glacier-permafrost interactions has been large.

Thirdly, a case study of the currently very small glacier, or glacieret, Omnsbreen which formed and largely disappeared during the LIA was studied in terms of glacial geomorphic evidence for permafrost interaction. Modelled permafrost distribution for the LIA suggest permafrost presence in the Omnsbreen surroundings during its formation and decay, and the landform assemblage present at Omnsbreen is considered representative for mountain glaciers terminating into permafrost. Permafrost is currently only present sporadically in the Omnsbreen surroundings, and the glacier and permafrost underwent a parallel disintegration.

The current geomorphic expression of Norway and northern Iceland is significantly affected by long-term interactions between the glacial, subglacial and ground thermal regimes.

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## Part II Papers

- Paper I: Lilleøren, K. S. and Etzelmüller, B. (2011) A regional inventory of rock glaciers and ice-cored moraines in Norway. *Geografiska Annaler, Series A - Physical Geography*, 93, 173-175-191.....81
- Paper II: Lilleøren, K. S., Gärtner-Roer, I., Etzelmüller, B., Käab, A., Westermann, S. and Guðmundsson, Á (*submitted manuscript*) Rock glaciers and ice-cored moraines in Tröllaskagi, northern Iceland – inventory and implications to Late-Pleistocene and Holocene landscape development. *Boreas* ..... 99
- Paper III: Lilleøren, K. S., Etzelmüller, B., Schuler, T. V., Gisnås, K. and Humlum, O. (2012) The relative age of mountain permafrost – estimation of Holocene permafrost limits in Norway. *Global and Planetary Change*, 92-93, 209-223..... 133
- Paper IV: Lilleøren, K. S., Humlum, O., Nesje, A. and Etzelmüller, B. (*submitted manuscript*) Holocene development and geomorphic processes at Omnsbreen, southern Norway – evidence for glacier-permafrost interactions. *The Holocene*..... 151



## Chapter 1

# Introduction, motivation and goals

The large-scale geomorphology of Norway has traditionally been described as consisting of two main components; (1) an old, 'paleic' surface and (2) a young, glacially sculptured landscape (Figure 1) (Reusch, 1901, Gjessing, 1967). The paleic surface is the generally high-elevated and lightly undulating vast areas that encompass the fjords and glacially eroded valleys. Edging most of the Norwegian coast, the low-relief areas which are referred to as 'the strandflat' resemble the paleic surface, but is interpreted to be both younger, and possibly formed by wave action in combination with a varying sea level during the Quaternary glaciations (Nansen, 1922). As such, the strandflat serve as a third important part of the large-scale geomorphology of Norway.

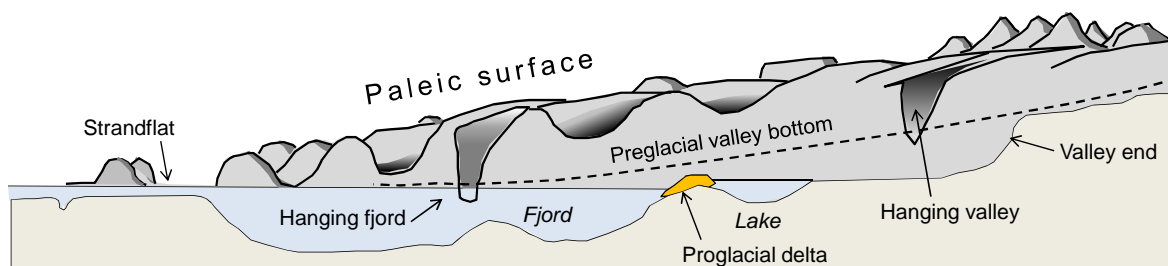


Figure 1 Elements of Norwegian large-scale geomorphology. Modified from J. Gjessing.

Processes working on the large-scale landscape form the middle- and small-scale geomorphology which constitutes a great variety of landforms. These are formed under glacial and periglacial conditions, and also by less climatically influenced processes, under

so-called 'azonal' conditions. At present, large regions of the Norwegian mainland are affected by processes related to cold climates, with mountain permafrost partly co-existing with glaciers, and seasonal and diurnal frost processes at lower altitudes. Norway is a country with large variations concerning climate and corresponding geomorphology, and the long coastline cause damped yearly temperature variations there compared to inland and mountain areas with more continental climate. Furthermore, the latitudinal effect lowers the different ecological zones in altitude from south towards north.

To varying degrees, the landmass of Norway is and has in the past been affected by what was recently termed 'cryoconditioned' processes by Berthling and Etzelmüller (2011). Both the periodic coverage of ice-sheets which characterize the Quaternary period, the periglacial environment of the outskirts and nunataks during glaciations, and alternating interglaciations are ultimately governed by thermally controlled factors. More specifically, large Quaternary periods were conditioned by temperatures below the freezing point of water. The concept of cryoconditioning in landscape evolution emphasize the importance of one common environmental characteristic, independent of glacial or periglacial conditions, namely that it is largely affected by the cryotic surface and subsurface thermal regime (Berthling and Etzelmüller, 2011).

When studying geomorphology, the correct classical methodology is to start with an inventory based on description and process tracing of the landforms in question, followed by investigations of the relationships between the involved processes, including observations and measurements. As a last step the development of today's landforms is put into a longer time perspective, where relevant time periods of the region's history is investigated in terms of landform formation (Ahnert, 1996). With this as a background, the main goal of my dissertation is to discuss landscape evolution and geomorphology on both small and large spatial and temporal scales as a product of the overall cryoconditioning of the landscape processes. This is done by (1) compiling inventories of intact and relict permafrost landforms of mainland Norway and northern Iceland, (2) modelling the Holocene permafrost variation of Norway, both spatially and temporally, (3) derive relative ages of the Norwegian mountain permafrost, and (4) investigations of glacier-permafrost interactions linked to the pro- and subglacial geomorphology of a small mountain glacier terminating in permafrost.

Part I of this thesis consists of a synthesized version of the four papers presented in Part II, and the combined relevance of the research is introduced and discussed. A thorough background is given on the morphogenesis of Fennoscandia, tentatively split between the large-scale landscape development of Fennoscandia, including orogeny, downwearing and

glaciations of the landmass, and meso- and small-scale landscape development. This includes processes working in periglacial and permafrost-affected environments, and the impact of glacier-permafrost interactions is introduced. The applied methods are elaborated, and shortly presented in the papers. The main results based on the papers of Part II are outlined, and discussed in a broader context. Here, the glacier-permafrost influence on landscape evolution is discussed both concerning the thermal regimes during glaciations and on the effect at smaller scales over the Holocene.

## Chapter 2

# Theoretical background

### 2.1 Concepts in geomorphology

Geomorphology is the study of the shape of the Earth, its surface processes and configuration of landforms (Anderson and Anderson, 2010), and the scientific branch has developed since the late 19<sup>th</sup> and early 20<sup>th</sup> centuries when first established by scientists like W. M. Davis, W. Penck and G. K. Gilbert. Geomorphology belongs to the earth sciences, and is often categorized as a sub-discipline of geography or geology. In geomorphology, the focus is process understanding and qualitative description of the physical environments, and how these are affected by the geographical position.

There are many concepts of importance when considering landscape development. In this section I will briefly introduce some, whereas others are left out. First, an environment in balance ('equilibrium') receives the same supply of matter as what is removed from an area unit; for example a part of a slope or a whole river catchment (Ahnert, 1994). The evolution of landforms is directly dependent on the amount of mass added to or removed from the system, and on the spatial and temporal variations in such a system. Isostatic rebound of the land mass after a glaciation (i.e. uplift) would count as a mass supply in this context. Further, the 'steady-state' system refers to a process-response system where the process rates are constant over time, while the 'dynamic equilibrium' refers to the relationship between the rates of the involved processes. In geomorphology it only makes sense to talk about open-system equilibriums, where the input of energy or matter keeps the processes active within

the system. By negative feedback, the process rates adjust themselves to that input, and have a tendency towards establishing a dynamic equilibrium between them.

In geomorphology, tendencies toward non-equilibrium occur when major changes in the input to the system happen. Major changes can for example be land uplift or major changes in climate, which for example leads to on-and-off systems of glaciations. In such cases, positive feedback effects guide the system away from an initial equilibrium state towards a period of non-equilibrium, and eventually towards a new state of equilibrium.

The relaxation of the system, the changes of the system components towards an equilibrium state, takes place rapidly at first, and then gradually more slowly and in smaller steps (Ballantyne, 2002). In addition, small variations will interrupt a long-term trend in nature, so determination of when the system has reached its equilibrium state may in nature be impossible. The adjustment towards the equilibrium is asymptotic, and in large systems several million years may have passed before the new state is reached (Summerfield, 1991). In such long time periods nothing in the system remains constant, but a large degree of equilibrium may have been reached long before, as most of the adjustment occur shortly after the new conditions on development were introduced (Peizhen et al., 2001). To establish equilibrium, new material needs to be added to the system continuously, and the observer needs to have an awareness of what kind of system that is investigated.

## **2.2 Landscape development**

### *2.2.1 Large-scale geomorphology*

Large-scale landforms, i.e. mountain chains, island arcs and ocean troughs are landforms formed mainly by endogenic processes, while meso- and small-scale landforms more often are the product of exogenic processes (Summerfield, 1991). The landform size is a product of the formation period (Figure 2), and all landmasses are a result of the combined work of upbuilding and downwearing processes. In this section a brief overview of the genesis of the Fennoscandia is given. The landmasses of the Scandinavian Peninsula (Norway and Sweden), Finland, Karelia and the Kola Peninsula are commonly termed Fennoscandia, and consist of Precambrian basement rocks formed by several orogeneses. At the onset of Cambrian, this landmass was denuded to a low-relief surface of which the remnants are referred to as the

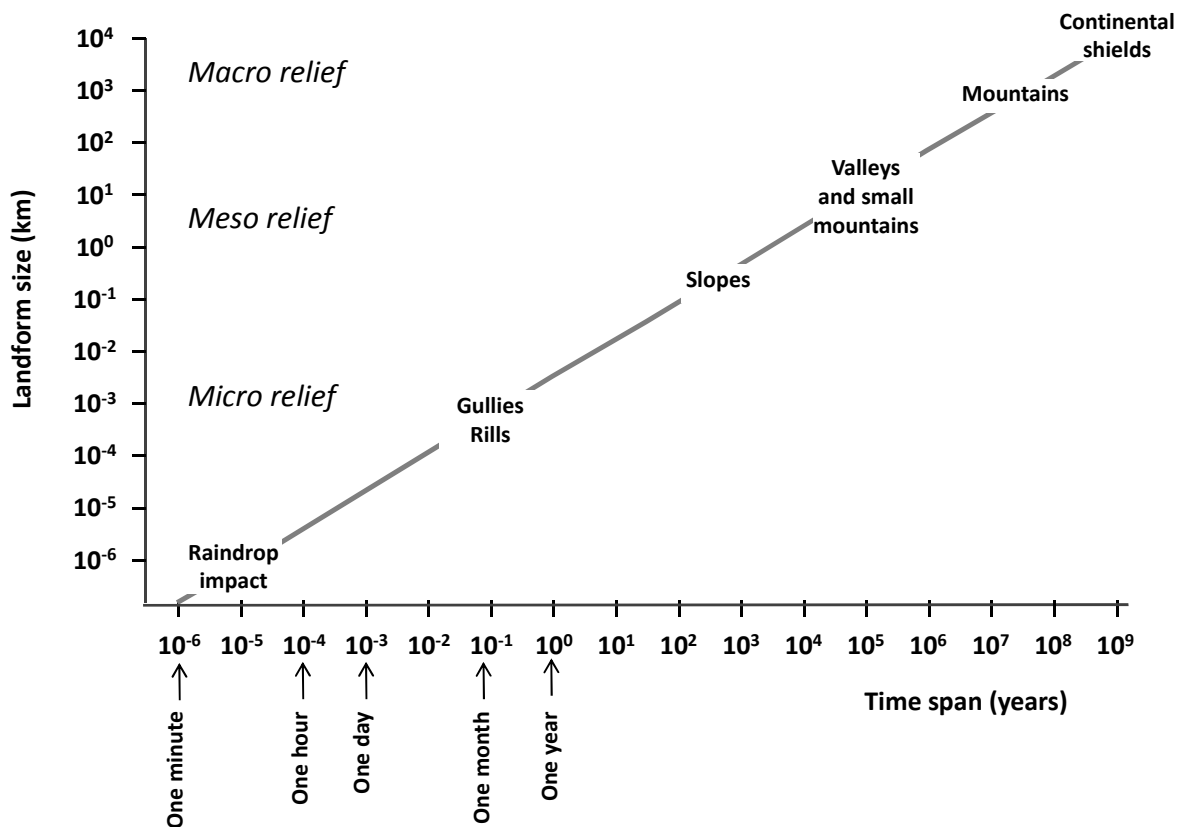


Figure 2 Relationship between size and duration of landforms, modified from Ahnert (1996).

sub-Cambrian Peneplain (Lidmar-Bergström and Näslund, 2005). The pre-Atlantic ocean Iapetus transgressed this peneplain, and marine sediments were deposited from Cambrian through Silurian (Högbom, 1910). Tectonic events caused the convergence of the two former continents Laurentia and Baltica during the early Ordovician (Dewey, 1969). As a result the Iapetus ocean narrowed and eventually closed, and the orogenesis of the Caledonides which currently are exposed continuously over more than 1,800 km in western Scandinavia occurred (Roberts and Gee, 1985). The Scandian phase of the Caledonian orogenesis is defined by the continent-continent part of the collision, and continued into Middle Devonian when the highest altitudes of up to 8-9 km were reached (Gabrielsen et al., 2010), and nappes were thrust eastwards over the Cambro-Silurian deposits (Gee and Sturt, 1985). This three-fold geology can be observed over large areas in Norway; the autochthonous Precambrian basement rocks below the deformed Cambro-Silurian marine deposits, and on top allochthonous basement thrustsheets. The Caledonides were strongly asymmetrical in appearance with a steep western flank, characterized by great relief (Gabrielsen et al., 2010). By the completion of the orogen the crust was over-thickened and several rift zones appeared,



causing the mountain range to collapse by crust extension (Fossen et al., 2008). At Late Carboniferous rifting of the new continent started, initiated at the Oslo graben areas. By Tertiary the main rift zone had shifted westwards and opened the Atlantic Ocean west of the remnants of the Caledonides, the Scandinavian mountain chain. As a result of this continental breakup, uplift occurred along the western flank of the now Fennoscandian land mass during Tertiary. Uplift following a continental breakup is generally asymmetrical along the passive margin, and as a result, the Fennoscandia is also at present steeper towards the ocean in west (Holtedahl, 1953, Lidmar-Bergström et al., 2000). At the onset of Tertiary, Fennoscandia underwent uplift in two central domes, the southern Scandes and the northern Scandes (Lidmar-Bergström and Näslund, 2005). Previously low-altitude plains were then elevated over several uplift events (Gabrielsen et al., 2010).

The characteristic decrease of global temperatures which characterizes the onset of the Quaternary led to glacier formation along the now elevated topography of the Fennoscandia. It is likely that the initial glaciations occurred along the main water divides, much like how the glacier distribution appears at present. During Quaternary, Fennoscandia was subjected to repeated glaciations, and glacial erosion commenced along the pre-existing fluvial valleys (Nesje and Whillans, 1994). Repeated cycles of glaciation of increasing duration led to the formation of fjords and U-shaped valleys which presently characterize the Fennoscandian landscape.

During interglacials, and surrounding the glaciers, periglacial and azonal processes worked. As both the Caledonides and the later Tertiary uplift had given the Fennoscandian landmass an asymmetrical appearance, erosion was largest in west due to the high gradient towards the sea. Both the erosion by Tertiary fluvial processes and later by glacial processes are structurally controlled (Lidmar-Bergström and Näslund, 2005), and the overall landscape asymmetry eventually shifted the main water divide eastwards. In between the incised and over-deepened glacier erosional zones, the Tertiary surfaces were kept largely intact and the glacial valleys end abruptly where they meet the paleic surfaces. These landscape features were early recognized by H. Reusch (1901), who following an excursion with the contemporary leading geomorphologist W. M. Davis ascribed the peneplains at different altitudes to belong to different stages in the geographical cycle. The 'geographical cycle' had short before been introduced by Davis (1899) and is a theoretical concept founded on James Hutton's principle of uniformitarianism. Davis argued that cycles where land uplift formed

the starting point was followed by mainly fluvial erosion and slope processes until an erosional baselevel was met as an end product. This baselevel was ultimately marked by sea level, unless the process was interrupted by a new uplift, and the rate of erosion slowed down as the landscape levelled, until the cycle was repeated (Figure 3).

The German geomorphologist W. Penck (1924) adopted and developed Davis' cyclic world view, but rejected the idea that the Earth's surface could be stable following an episode of rapid uplift. His alternative view was that the rate of slope retreat and valley incision is related to the tectonic state of an area, either as increasing or decreasing uplifts, or as a stagnated landmass. This view implies that the Earth's shape is a result of reactions between opposite forces, where erosion in one place led to deposition somewhere else, and the concept of a 'dynamic equilibrium' was introduced. In Penck's dynamic equilibrium, the overall landscape slowly change while within a given time slot the sediment input to a geomorphic system is equal to what is evacuated from it.

While both Davis and Penck thought of the slope development as a gradually decrease in angle, until a peneplain were reached, L. C. King (1962) introduced the concept of parallel slope retreat. Like Davis, King (1953) too imagined the landscape development as being cyclic, where long periods of tectonic standstill were separated by rapid events, implying

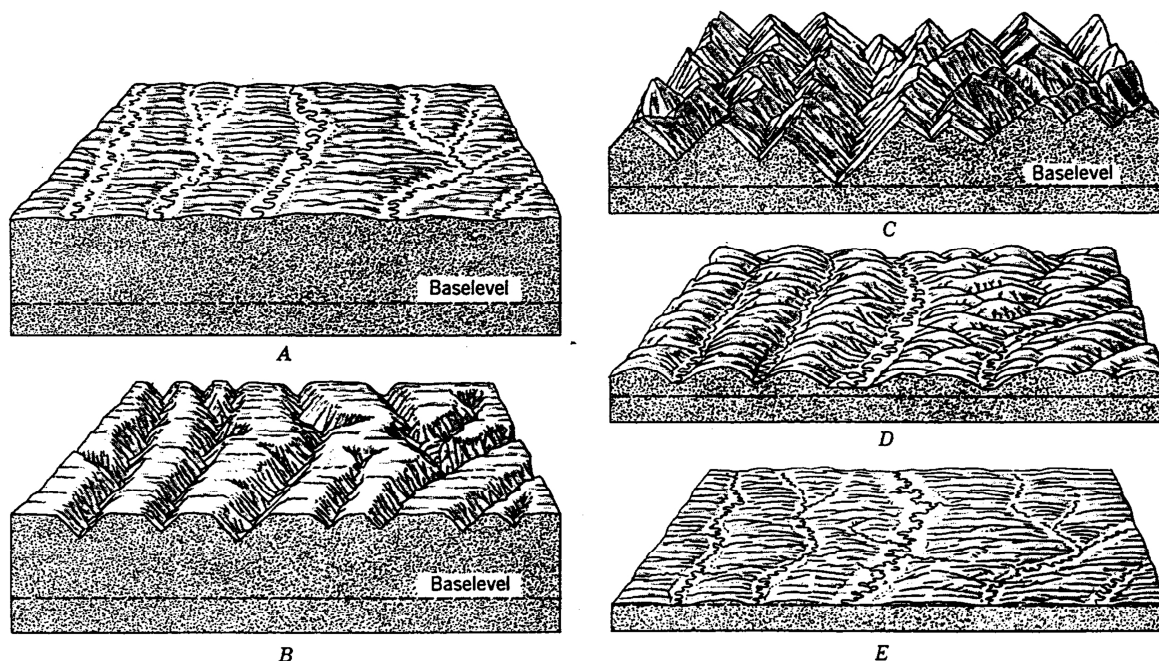


Figure 3 Illustration of W.M.Davis' 'geographical cycle' from the uplift of the landmass (A) via fluvial erosion and gradual downwearing until the erosional end product, the 'baselevel' is reached (E).

discontinuous uplift. Here, erosion occurred inwards from the coast, resulting in an age differentiation within what he termed pediplains; they are multichronous. Remnants between the pediplains were termed inselbergs. The main arguments against all these landscape development theories from contemporary scientists was that the cyclic views were theoretical and idealized, and that the importance of climate variability in landform development was undervalued (Summerfield, 1991).

The importance of climatic variability in geomorphologic processes was emphasized by J. Büdel (1963). Büdel (1982) pointed out that his predecessors did not base their systems on morphological criteria, i.e. which forms would develop within specific climate zones, but rather emphasized that surface processes specializes within different climatic zones. He introduced his peneplain concept based on etching surfaces and double planation where stairways with escarpments will develop on the flanks of crustal uplifts and around etched intra-mountain basins. Based on experience from the arid tropics, Büdel suggested chemical weathering to occur along two important zones; along the surface-air interface and along the sediment-bedrock interface. Etch-driven lowering of the landscape could continue on plains until uplifted above the normally tropical or sub-tropical climate feasible for double planations to work effectively. This process further implies that all parts of one plain have the same age; they are monochronous. According to Büdel, the Scandinavian elevated peneplains are the result of deep weathering and etching during the warm and humid Tertiary period, whereas erosional landforms like *roche moutonnées* were irregularities in the bedrock where weathering products were removed by the glaciers and successively smoothed.

In southern Norway, the mountain regions of Dovre, Rondane and Jotunheimen reach 1200-1500 m above the lowest altitude of the paleic surface. Except for Jotunheimen and the northwestern Romsdalen in southern Norway (Figure 4), which have an alpine character, mountain areas are generally only moderately affected by glacial erosion, and the pre-Quaternary appearance kept intact (Lidmar-Bergström and Näslund, 2005, Etzelmüller et al., 2007b).

### 2.2.2 *Paraglacial processes*

The effect of repeated glaciations in terms of landscape reworking of an area is well known, and glacial erosion is often referred to as the single most effective erosive agent on Earth. The presence of a glaciation introduces a new regime of erosional and depositional processes,



*Figure 4 Key map of Norway, where places mentioned in the thesis are marked.*

and both magnitude, duration and the extent of forces are different from those working in a ‘normal’, subaerial environment (Church and Ryder, 1972). Less acknowledged are the erosional processes working between periods of glaciations, in interglaciations or in interstadials. Formerly glaciated areas are subject to massive reworking of sediments and a change in processes once the glacier retreats. These processes are commonly termed

‘paraglacial processes’, i.e. processes involving formerly glaciated areas, going on in glacial deposits, and/or are a direct consequence of the glacier’s work in the area.

Church and Ryder (1972) first defined the term ‘paraglacial’ as “nonglacial processes that are directly conditioned by glaciation”, and later specified that paraglacial processes are not restricted to the late Pleistocene or to other “closing phases” of a glaciation, but are a continuing characteristic of mountain walls (Church and Ryder, 1989). Later, Ballantyne (2002) defined paraglacial geomorphology as glacially conditioned sediment availability, implying that once a glacier retreats from an area, the exposed environment is likely to undergo fast changes. Thus, the term ‘paraglacial’ refers to processes at work in a transitional period between two stages of landscape equilibrium (Etzelmüller and Frauenfelder, 2009).

Among other processes within the paraglacial framework, over-steepened glacial rock walls may result in slope failure or enhanced rockfall activity if not in balance with the bedrock’s mass strength, slopes consisting of unconsolidated material are more vulnerable to reworking by debris flows, snow avalanches and slope wash, glacier forelands are more exposed to wind erosion and frost action, and rivers entrain and redistribute large amounts of unconsolidated sediments of glacial origin. Consequently, a somewhat radical theory has been forwarded stating that most of the valley widening effect of numerous glaciations are happening in between, and not during, glaciations (Jarman, 2009). According to this interpretation, most rock slope failures are occurring immediately after a glaciation when periglacial processes take over, leading to talus build-up, solifluction, gelifluction, landslides and so on in unconsolidated material, while freeze-thaw cycles in bedrock cause fractioning and rock fall events. An advancing glacier will remove the sediment deposited during the ice-free period, polish and induce stress on the valley slopes, eventually retreat and the paraglacial cycle starts over again.

In the periods immediately following a glaciation, the stage is set perfectly to create new periglacial landforms, as the area is adjusting from a glacial to a new non-glacial equilibrium. Here, the paraglacial framework provides a link between the glacial and the periglacial processes (Etzelmüller and Frauenfelder, 2009). Both mass movement due to stress release and tectonic events due to land uplift will generate sediment availability in slopes affected by glaciation. In such slopes, a continuum of landforms from rock glaciers, which require permafrost, to solifluction lobes, which require seasonal freezing, will form. Following a strict definition of the term ‘paraglacial’, isostatic rebound events subsequent to

glacier retreat are not a “*paraglacial process as it is an indirect tectonic response rather than a process operating at the earth’s surface*” (Ballantyne, 2002: p. 1938). Still, such events provide sediment for periglacial processes to rework. The rate of which the adjustment from one equilibrium state to the next occurs is highly variable, from slow processes that involve large volumes of mass, to rapid processes involving small to medium volumes of mass (high-magnitude – low-frequency vs. low-magnitude – high-frequency). The paraglacial processes as a framework is useful and should be appreciated as a valuable principle when discussing the Holocene landscape development.

### **2.3 Periglacial geomorphology and permafrost**

The term ‘periglacial’ was introduced for the first time by Lozinski (1909, 1912) to denote geomorphic and climatic processes that worked in the outskirts (in the *periphery*) of the Fennoscandian ice-sheet. Following this first introduction the understanding of the term has shifted towards a general description of processes or geomorphology occurring in areas dominated by freeze-thaw cycles. Despite efforts to clarify and suggest rigorous definitions (e.g. French and Thorn, 2006), ambiguity remains (Berthling and Etzelmüller, 2011). While some scientists require perennially frozen ground, i.e. permafrost, to be present in the periglacial environment (e.g. Péwé, 1969), for example French (2007) summarizes the current understanding of ‘periglacial geomorphology’ as a sub-discipline within geomorphology concerned with cold, non-glacial landforms, and do not mention permafrost as a criteria. An alternative term to denote geomorphology of cold regions in general was introduced by Berthling and Etzelmüller (2011), namely ‘cryogeomorphology’, a spin-off term from the concept of ‘cryoconditioned’ landscapes. The advantage of this notion is the unification of glacial and periglacial processes on the basis of belonging to a certain climate type and includes both periglacial and glacial geomorphology. As such, two disciplines within geomorphology which are often treated separately in the scientific community can be joined, at least concerning nomenclature. The permafrost in the Nordic countries belongs to the periglacial zone, i.e. areas outside glaciers and north of or above the timber line, but this is not always the case. Permafrost also underlay forested areas of Mongolia, Alaska and Canada (French, 2007).

By the International Permafrost Association (IPA), permafrost is defined as ‘soil or rock with included ice and organic debris that remains at or below 0 °C for at least two consecutive years’. Hence, glacier ice is excluded from the permafrost regime (IPA, 1988).

The surface layer which thaws every summer is also excluded from the permafrost regime, and is termed the *active layer*. Since permafrost is defined by temperature alone, the latter definition is not accurate, as unfrozen water can exist at subzero temperatures, thus includes the upper part of the permafrost in the active layer (French, 2007). The thickness of the active layer varies largely due to differences of heat conductivity associated with bedrock types, surface cover, ice and/or water content in the substrata, and of course due to climate characteristics, for example as continental versus maritime climate.

A number of landforms exclusively forms and exists in permafrost environments, and based on the state of activity of such landforms they can be used as a proxy of present and former permafrost occurrence. In Scandinavia, four types of landforms indicate the presence of permafrost; palsas (Sollid and Sørbel, 1974, Seppälä, 1994, Sollid and Sørbel, 1998), rock glaciers (Shakesby et al., 1987, Sollid and Sørbel, 1992), ice-cored moraines (Østrem, 1964, King, 1986), and ice-wedge polygons (Svensson, 1962, Sollid et al., 1973, Svensson, 1992). With the exception of ice-wedge polygons, active examples of the remaining three types of landforms are known to currently exist in Norway. Palsas are a special form of permafrost landforms since these develop and exist in bogs where the thermal properties of peat are decisive of their existence rather than a specific set of climatic conditions. Therefore, they exist in a zone stretching south of or lower than the regional permafrost belt, and are also delimited northwards and in altitude by climatic restrictions on peat production.

The perhaps most striking geomorphic expression of permafrost is rock glaciers. Rock glaciers are glacier-like landforms consisting of unconsolidated material of most grain-sizes, which creep downslope due to deformation of the interstitial ice and gravitational pull (Barsch, 1996, Haeberli et al., 2006).

According to Barsch (1996) several geomorphological conditions need to be met in order for the landform to be termed 'rock glacier' ('rockglacier' in Barsch, 1996); (1) a rock glacier stand above the adjacent terrain (typically 10-20 m), (2) they have steep front and side slopes which appear light due to exposure of unweathered debris, (3) compared to the front and side slopes the upper surface has a gentle slope, and (4) in the upper end the landform either grade into a talus slope or a depression exist between the rock glacier and the headwall which now or in the past has been occupied by an ice-field or a small glacier (Figure 5). Due to local topography several exceptions from this set of rules exist (Barsch, 1996). In literature, a long-lasting controversy exists concerning nomenclature and definitions of rock

glaciers, where especially two different scientific positions are debated: (1) that rock glaciers are the geomorphic expression of slow creep in permafrost ground and (2) that rock glaciers and rock glacier-like landforms are the visible expressions of a multitude of processes, respectively termed the ‘permafrost creep school’ and the ‘continuum school’ by Berthling (2011). Very briefly summarized the permafrost creep school, which by far has done most of the quantitative research in the area, is based on the assumption that rock glaciers are the only true geomorphic expression of creep of unconsolidated material in a steady-state permafrost environment, which can occur in material such as talus slopes, till and moraine-deposits (Haeberli, 1985, Barsch, 1996). This group do emphasize that ground ice can arise from a variety of processes, including the burial of snow and ice (Haeberli, 2000). The continuum school thus takes a continuity stand, claiming that all landforms that resemble rock glaciers in morphology should be termed rock glaciers, such as for example heavily debris-covered glaciers and creeping ice-cored moraines, and follow the morphological definitions of Capps (1910) rather than genesis and processes (e.g. Martin and Whalley, 1987b, Whalley and Martin, 1992, Clark et al., 1994, Hamilton and Whalley, 1995). For the latter group, permafrost is not specified as a prerequisite for a rock glacier to form and maintain dynamics independent of for example the glacier upslope, thus creating new controversies. In Iceland, a permafrost and rock glacier controversy originate from this latter group, when the Nautárdalur rock glacier in Tröllaskagi was classified as a ‘glacier ice-cored rock glacier’ terminating in a non-permafrost environment as stated by Martin and Whalley (1987a), and is discussed in section 2.3.2.

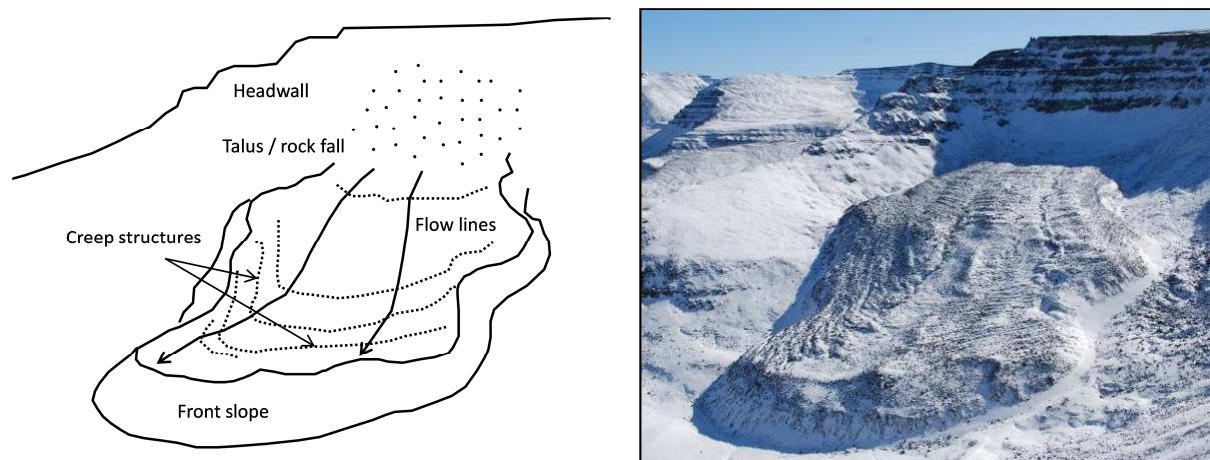


Figure 5 Characteristics of an active talus-derived rock glacier, Tröllaskagi, Iceland.



Another aspect of the rock glacier classification scheme is whether the landform should be identified mainly on processes and genesis ('the permafrost creep school' following Berthling, 2011, Haeberli, 1985, Barsch, 1996), or that a morphological unity is sufficient ('the continuum school', e.g. Capps, 1910, Martin and Whalley, 1987b). Again, others entirely leave the permafrost environment stating that rock glaciers can form when the debris-cover on glaciers overcome a critical thickness or that landslides or large rock fall events eventually can develop creep movement (Whalley and Martin, 1992). However, both debris-covered glaciers and original landslide deposits which develop secondary creep can turn into rock glaciers if they exist in a permafrost environment. For example, the rock glaciers in the alpine regions of especially northern Norway is interpreted to have been initiated as landslide events during the deglaciation as a paraglacial response, which eventually developed secondary creep (Tolgensbakk and Sollid, 1988). A different example is the currently inactive rock glacier known as the 'Verkilsdalen landslide' in Rondane in southern Norway (Barsch and Treter, 1976, Dawson et al., 1986).

Stable ice-cored moraines are found in regions where cold or polythermal glaciers terminate in permafrost environments. As a general landform, this type of moraine is not exclusively formed in permafrost environments. However, in non-permafrost regions these ice-cored structures are prone to rapid mass wasting, and are not stable over more than a few decades (e.g. Driscoll, 1980, Krüger and Kjær, 2000). Conversely, within a permafrost environment, the ice-core may survive millennia if the debris cover on the glacier front is thicker than the active layer depth, causing protection from ice degradation (Etzelmüller and Hagen, 2005). In those cases, ice-cored moraines become indicators of permafrost presence, as they by definition can only be stable in permafrost environments. In this way, both rock glaciers and ice-cored moraines serve as climate indicators as they reflect a certain ground thermal regime.

### *2.3.1 Permafrost in Norway*

In Norway, permafrost is a widespread phenomenon (Figure 6), which currently underlie approximately three times the area covered by glaciers, but is obviously harder to observe given its definition as a subsurface and strictly thermally defined phenomenon (King, 1983, King, 1986, Ødegård et al., 1996). However, due to initiatives such as the PACE (Permafrost and Climate in Europe) project (Harris and Vonder Muhll, 2001), TSP (Thermal State of

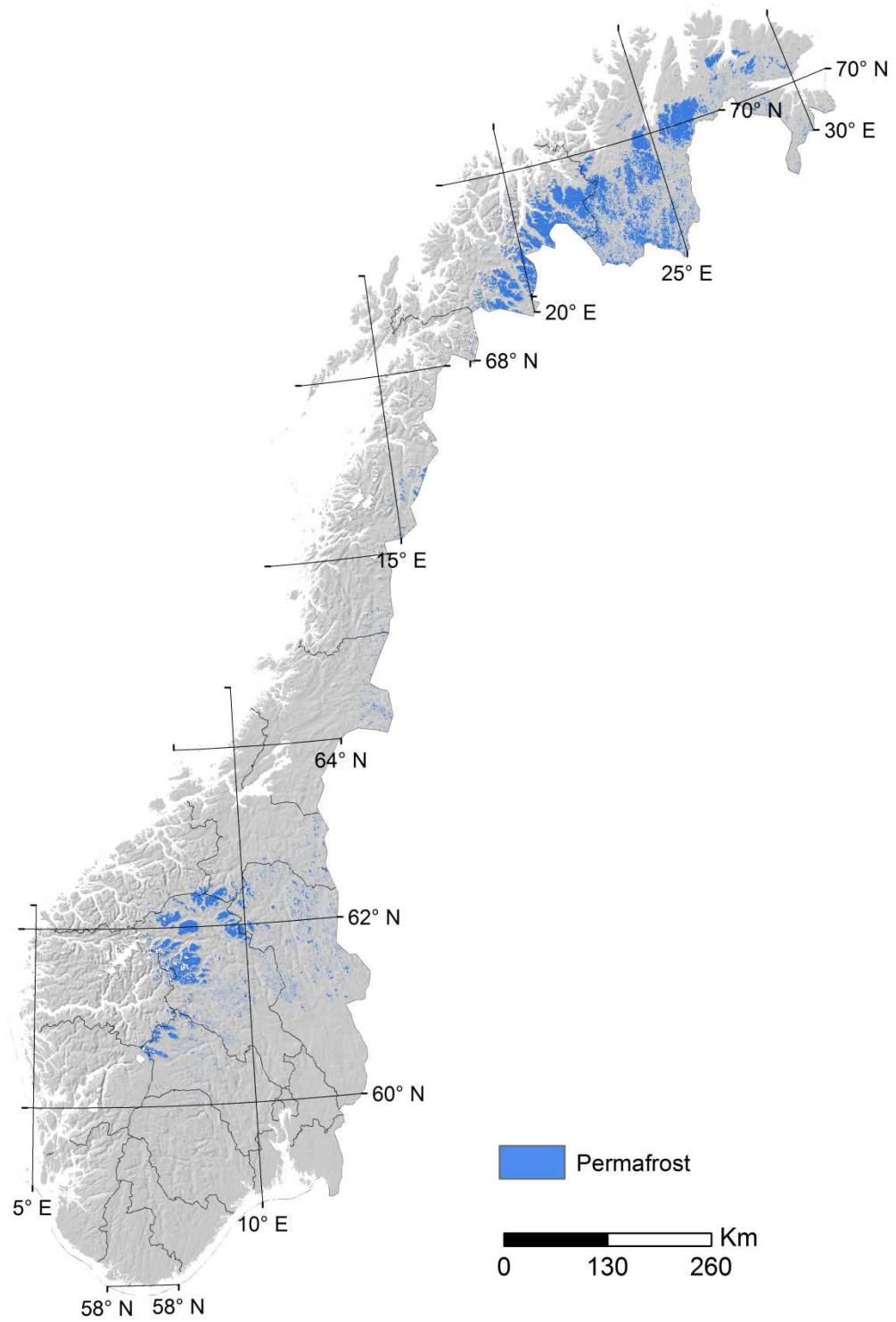


Figure 6 Permafrost distribution in Norway, as modelled by Gishnås (2011).

Permafrost) (Christiansen et al., 2010) and the CryoLINK project (Etzelmüller et al., 2009) the current knowledge of permafrost in Scandinavia has been greatly improved over the last decade. Several boreholes in permafrost regions have been drilled and provide direct observations of permafrost temperature and thermal state while series of boreholes in altitudinal transects as well as in climatic transects lead to increased understanding of permafrost zonation. Such direct observations have made it possible to calibrate 1D and 2D permafrost models (Gisnås, 2011, Hipp et al., 2012). As a regional pattern in Scandinavia, the lower limits of mountain permafrost altitude (MPA) decrease in a transect from west towards east, where the lower permafrost limit of e.g. Jotunheimen and Dovrefjell is c. 1550 m a.s.l. (Ødegård et al., 1996, Isaksen et al., 2002). The distribution of permafrost is highly dependent on topographic effects such as slope and aspect, surface characteristics and local snow conditions, where for example locally on Dovrefjell on particular snow-blown sites, permafrost is observed at 1350 m a.s.l. (Sollid et al., 2003). Permafrost probably also exists west of these mentioned areas at summits exceeding c. 1600 m a.s.l., while in eastern parts of southern Norway the regional lower limit is probably around 1300 m a.s.l. (Etzelmüller et al., 2003, Heggem et al., 2005). The observational basis in these areas are however limited.

### 2.3.2 *Permafrost in Iceland – the permafrost-rock glacier controversy*

In the official International Permafrost Association (IPA) map the Icelandic permafrost distribution is restricted to palsa areas in central Iceland, covering an area of c. 180 km<sup>2</sup> at altitudes between 460 and 720 m a.s.l. (Priesnitz and Schunke, 1978, Brown et al., 1995). However, ground temperature monitoring in boreholes indicate a total permafrost area in Iceland of c. 7000-8000 km<sup>2</sup>, and a decline of the lower limit of permafrost from southeast towards northwest (Etzelmüller et al., 2007a, Farbrot et al., 2007b). Permafrost is most common in the mountainous areas of the Tröllaskagi peninsula, around the Askja crater, at the northern and southern margins of Sprengisandur and in northeast Iceland at Smjörfjöll (Figure 7) (Etzelmüller et al., 2007a).

The existence of both intact and relict rock glaciers in Iceland has been a subject of discussion since the 1980s when Martin and Whalley (1987a) published a paper on investigations of one specific rock glacier in Tröllaskagi, followed by a later publication on a wider study sample (Whalley and Martin, 1994), and at the same time denied permafrost as a

prerequisite of rock glacier formation in this area. The controversy continued when a master thesis inventorying rock glaciers in Iceland was

submitted (Guðmundsson, 2000), presenting a systematic compilation of active, inactive and relict rock glaciers. Together, these publications formed a renewed and important discussion on rock glaciers and permafrost.

In the first-mentioned publications above, rock glaciers in the Tröllaskagi area were investigated as strictly glacier-related landforms, and the most thoroughly studied rock glacier was termed a ‘glacier ice-cored rock glacier’ in the paper (Whalley and Martin, 1994). This hypothesis was based on the observation that rock glaciers in Iceland most often are associated with small glaciers (<1 km<sup>2</sup>) and that ice with debris bands often is visible in the lower parts of glaciers, feeding debris to the rock glacier downslope of the glacier (Whalley et al., 1995b). By analysing mean annual air temperatures (MAAT) from Akureyri, as a base for calculated lapse rates for mountain sites, Whalley and Martin (1994) concluded that an MAAT of -1.5 °C at the rock glacier snouts make permafrost presence unlikely. Regarding the formation (genesis) of these rock glaciers, Whalley and Martin (1994) assumed that the ‘Little Ice Age’ (LIA) was most likely the formation period, since debris transport rates was higher then. Thus, they have an age of approximately 200 years.

In the last-mentioned publication above (Guðmundsson, 2000), a high number of large landforms consisting of unconsolidated material and often, but not always, situated close to sea level were classified as relict rock glaciers, in addition to less controversial active rock glaciers at higher altitude. These low-lying landforms sometimes resemble relict rock glaciers, as they are clearly restricted spatially, have uneven surfaces, creep structures and hilly local topography, but are many times larger than relict rock glaciers familiar from other places in the world. These landforms are commonly interpreted as landslides in Iceland (Thorarinsson, 1954, Jónsson, 1976, Whalley et al., 1983) since they normally have a clearly defined source area and are most often referred to as caused by stress-release following the deglaciation of Iceland, but also this classification is debated. However, the idea of relict rock glaciers at sea level was at the time of the publication not in accordance with the existing deglaciation model of Iceland, and there was simply not enough time in this model to develop such large rock glaciers only by slow creep processes during and/or immediately following the deglaciation.

More recently, additional studies of rock glaciers in Iceland have been carried out by several researchers especially in the surroundings of Hólar in Hjaltadalur (e.g. Wangensteen et al., 2006, Farbrot et al., 2007a, Kellerer-Pirklbauer et al., 2008). Here, both active and

relict rock glaciers exist, and have been studied in terms of distribution, surface displacement rates and relative age determination. First, these papers all state that the rock glaciers observed in the area are in fact the geomorphic expression of long-term cryogenic processes in the ground like those familiar from other parts of the periglacial areas of the Earth. Second, they all acknowledge the presence of permafrost as a present and widespread phenomenon in high-altitude areas of Iceland, and as a former phenomenon in low-altitude areas, based on for example permafrost modelling and temperature observations from boreholes (Etzelmüller et al., 2007a, Farbrót et al., 2007a).

As a general observation, the rock glaciers in Iceland are very well developed landforms, and represent a wide variety of landforms, both concerning genesis and activity. The Tröllaskagi region is characterized by extensive local glaciation, and the general potential of glacier-permafrost interactions is high, with a substantial number of ice-cored moraines as a result.

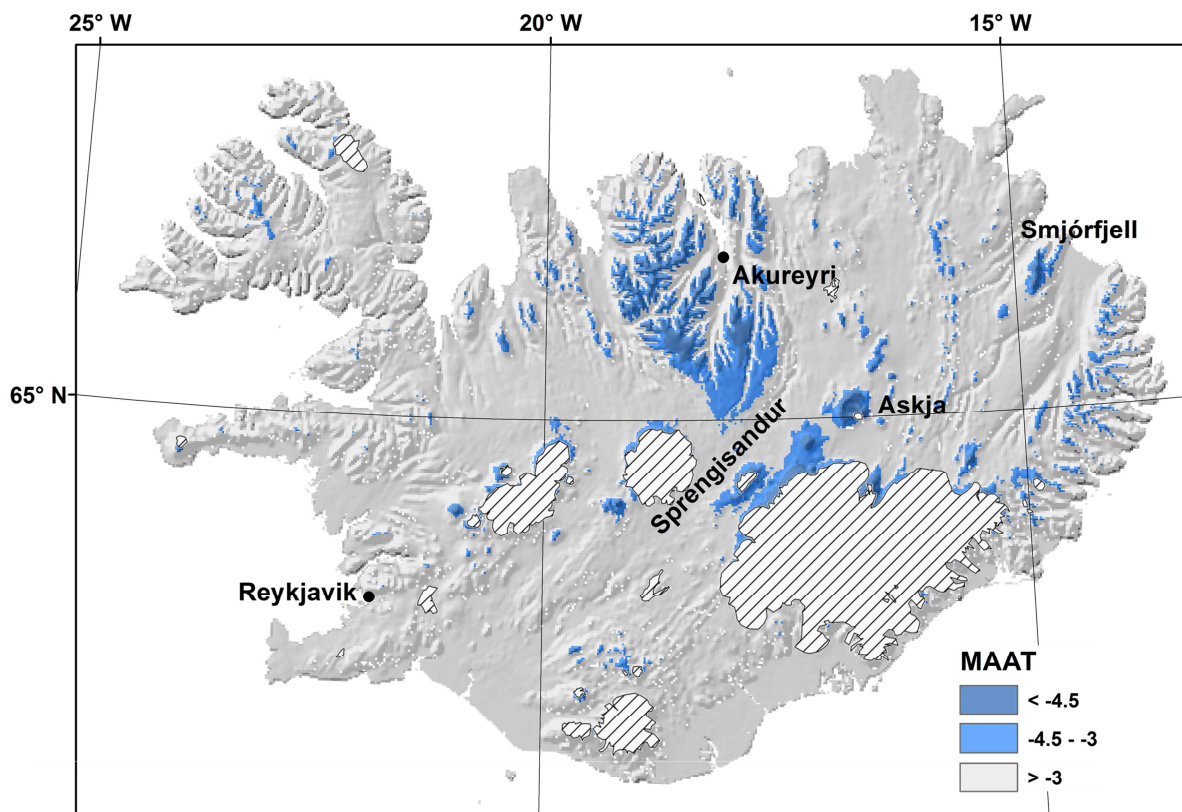


Figure 7 Predicted permafrost extent in Iceland, based on mean annual air temperatures (MAAT). Continuous permafrost may occur at  $-4.5^{\circ}\text{C}$ , discontinuous permafrost between  $-4.5$  and  $3^{\circ}\text{C}$ , and at temperatures higher than  $-3^{\circ}\text{C}$ , permafrost is unlikely.

## 2.4 Glacier-permafrost interactions

Shumskii (1964) defined glaciers as part of the hydrosphere, and permafrost as part the lithosphere. These definitions have persisted in the scientific community and are still valid. This is partly based on the assumption that glaciers and permafrost mutually exclude the other component, as substantial thicknesses of glacier ice isolate the ground thermal regime from the atmosphere, and that heat-generating subglacial processes in a polythermal glacier weaken the underlying permafrost. Further, where the glacier is cold-based, preservation of the underlying landscape is expected and such glaciers are considered ineffective in terms of geomorphology. This divergence of research fields have led to increasingly disconnected scientific research societies, which only to a limited degree cooperate or use consistent terms (Spagnolo et al., 2012). Recently, efforts have been made to unite these research fields, for example in publications like Harris and Murton (2005a) and Waller et al. (2012).

Clear geomorphic significance and specific landform assemblages are connected to areas prone to the interfingering processes between glaciers and permafrost (Harris and Murton, 2005b). Recently, the concept of ‘cryoconditioned’ landscapes was introduced to describe “*the interaction of cryotic surface and subsurface thermal regimes and geomorphic processes*” (Berthling and Etzelmüller, 2011, p. 380), and stresses the interconnectivity between glacial, periglacial and azonal processes at work in cold-climate environments.

The original meaning of the word ‘periglacial’ included the association of frost action in permafrost and proglacial environments (Lozinski, 1909). Where permafrost is most extensive, that is in polar or mountain regions, glaciers also tend to build up, and currently ice-free areas surrounding the ice sheets of Greenland and Antarctica are characterized by extensive permafrost. Little is known about the subglacial regime of the present-day ice-sheets due to the inaccessibility of such areas. Direct observations on ice-sheet temperature exist from boreholes, and at ice-divides where little lateral glacier flow occur, cold ice is observed deep into the ice (Dyke, 1993). In former subglacial environments, evidence of glacier-permafrost interactions exists. For example, studies have shown that the extent of glacier-permafrost interactions affects the largest areas during periods of ice-sheet advances, when permafrost in the glacial periphery was overridden (Mathews and Mackay, 1959, Cutler et al., 2000). Also, the glaciers tend to be cold in the first phases when they form in regions that have cold climate and thick permafrost (Dyke, 1993). During the Pleistocene glaciations, permafrost developed widely around the margins of the ice-sheets (Vandenberghe and Pissart, 1993, Ballantyne and Harris, 1997).

Like the large ice-sheets, also smaller glaciers provide a thermal offset between the air temperature and the ground thermal regime in permafrost regions. This means that glaciers terminating in permafrost are either partly or completely cold, that is they are either polythermal or cold glaciers, respectively. How large the sub-zero mass of the glacier is depends on factors like winter air temperatures, glacier dynamics and snow cover over summer (Liestøl, 2000), in addition to the underlying permafrost thickness (Dyke, 1993). Polythermal glaciers are frozen to the ground at least in marginal zones where the ice is thinnest, while ice in the accumulation zone is warmed due to meltwater penetration and latent heat release in the firn in regions where summer melt occurs (Paterson, 1994). Independent of the climatic situation also small, initially temperate glaciers will turn cold or polythermal during retreat. In such situations the ablation zone increases leading more of the meltwater runoff to evacuate the glacial system and less energy release occurs in the reduced firn zone (Paterson, 1994, Hock, 2003). Locally, permafrost will develop in connection to such retreating glaciers.

Geomorphologically, glacier-permafrost interactions are imprinted in the presence of landforms like open-system pingos, moraine-derived rock glaciers, ice-cored and push moraines (Boulton, 1972, Liestøl, 1977, Benn and Evans, 1998, Lyså and Lønne, 2001, Etzelmüller and Hagen, 2005). In a more indirect way, the presence of paleic surfaces in large parts of high-altitude Scandinavia, often covered by *in situ* blockfields, can be interpreted to indicate the presence of non-erosive cold-based ice-sheets and thus a geomorphic feature of glacier-permafrost interactions (Berthling and Etzelmüller, 2011). Conflicting views of the age and implications for ice-sheet dimensions represented by blockfields exist in the scientific literature, and they have traditionally been interpreted as either palaeo-nunatak phenomena (e.g. Nesje et al., 1987, Nesje, 1989, Nesje and Dahl, 1990, Brook et al., 1996) or as protected and preserved beneath a cold-based ice (e.g. Follestad, 1990, Kleman, 1994, Sollid and Sørbel, 1994, Kleman and Hättestrand, 1999, Fjellanger et al., 2006). Later the view on age and significance of blockfields has been modified to include both cases at different temporal and spatial scales (Ballantyne, 1998, Goodfellow, 2007, Ballantyne, 2010). For example, cosmogenic datings suggest ice-free conditions in the alpine areas of western southern Norway since >55,000 years BP (Brook et al., 1996, Goehring et al., 2008) while in other areas evidence of blockfields overridden by glaciers exist (Rea et al., 1996, Fjellanger et al., 2006, Goehring et al., 2008). However, cosmogenic datings of tor emergence from

blockfields suggest Middle Pleistocene ages (Phillips et al., 2006b, Darmody et al., 2008), implying that the surrounding blockfields was subsequently lowered (Ballantyne, 2010). An updated blockfield formation model which accounts for these observations was therefore recently presented by Ballantyne (2010). In this model a near-horizontal preglacial surface covered by a few metres of saprolite is assumed as the starting point, and a Quaternary surface lowering of several metres occurred. First, pre-Pleistocene chemical weathering works along a horizontal weathering front parallel to the surface, gradually transforming to bedrock in depth via a zone of pronounced corestones (Roaldset et al., 1982, Ballantyne, 2010). Entering the Pleistocene, frost weathering gradually takes over the rock jointing process as the residual saprolite is stripped from the surface and the remaining corestones produce a 'proto-blockfield' (Ballantyne, 2010). Over the Pleistocene, periglacial processes accentuate the blockfields by selective vertical frost sorting processes lifting the large blocks, which are altered corestones and/or frost-wedged rocks, towards the surface, whereas gradually finer material is situated below and within the bedrock joints (Ballantyne, 2010).

The Quaternary mass removal from the blockfields is explained by episodic events of glacial erosion by either rapid temperature changes in the ice during or at the end of glaciations (Dredge, 2000, André, 2004, Hall and Phillips, 2006), or by sediment entrainment and deformation by cold-based ice (Fitzsimons et al., 1999, Cuffey et al., 2000, Waller et al., 2012). Lowering of blockfields where no evidence of glacial overriding exists is harder to explain, but slow lateral mass movement (Small et al., 1999, Anderson, 2002), plug-like flow over cold permafrost in low-gradient slopes (Egginton and French, 1985, Lewkowicz and Clarke, 1998, Matsuoka, 2001), and aeolian or fluvial removal of fine-grained products of superficial clast weathering (Ballantyne and Harris, 1997, André, 2002, Hall and André, 2003, Matsuoka and Murton, 2008) have been suggested (Ballantyne, 2010). Lastly, permafrost presence in periods of active blockfield development is assumed in this model, causing drainage limitation during freeze-back and effective frost wedging. Thus, the blockfield depth equals the active layer thickness (Ballantyne, 2010).

The openwork structure of the Scandinavian blockfields provides a negative ground thermal anomaly compared to adjacent finer-grained surface cover or bedrock (Harris and Pedersen, 1998, Gorbunov et al., 2004, Juliussen and Humlum, 2007b, Juliussen and Humlum, 2007a). Therefore, blockfields are prone to maintain permafrost in areas at marginal permafrost conditions which are otherwise permafrost free, and Berthling and Etzelmüller (2011) propose positive feedback processes between blockfields, permafrost and



cold-based glacier cover. Glaciers aggrading into areas of extensive and thick permafrost will attain cold-based conditions until substantial thicknesses are achieved (Dyke, 1993).

For areas affected by the Fennoscandian glaciation, several zones based on assemblages of landforms interpreted to belonging to different subglacial thermal regimes have been proposed by Sollid and Sørbel (1994). Here, one central zone is characterized by landforms such as lateral meltwater channels, Rogen (ribbed) moraines and drumlinoid landforms located in surface depressions (Sollid and Sørbel, 1994). The areas bordering this zone appear quite different, with characteristic landforms such as eskers, drumlinoid landforms and some glacier-marginal deposits. These authors interpret the different geomorphological characteristics as marking a gradual change of the thermal regime of the ice-sheet from cold-based to warm-based outwards from the culmination zones. Hättestrand and Kleman (1999), however, associate the formation of Rogen moraines itself to depressions in the landscape within the cold-based ice regions, whereas Sollid and Sørbel (1984) ascribe transition between physical properties of the ice to the formation of this landform. In sum, several studies indicate large areas of the Scandinavian peninsula to have been covered by cold-based ice at least partly during the glaciations; eastern parts of Norway, most of Sweden north of 61°N, northern parts of the Gulf of Bothnia and also north-eastern parts of Finland (Lundqvist, 1989, Kleman, 1992, Kleman et al., 1992, Kleman and Borgström, 1994, Hättestrand, 1997, Kleman and Hättestrand, 1999).

## Chapter 3

# Methods

Geomorphology has traditionally been a qualitative discipline within science, where map products made via field and/or air photo interpretations have served as both research tools and as results. At present, the discipline moves towards more quantitative approaches, where especially the increased computational power and easily available high-resolution web-based aerial photos are valuable research tools. This thesis takes advantage of this development, but it is still based on qualitative approaches.

### 3.1 Landform inventories, Norway and Iceland

#### 3.1.1 *Compilation*

An important and natural first step in geomorphology is to establish databases or inventories of landforms and landscape components (Ahnert, 1996).

For mainland Norway and Iceland permafrost landform inventories were compiled based on interpretation of aerial imagery. In these cases permafrost landforms include rock glaciers, either talus-derived or moraine-derived, and large stable ice-cored moraines. Azonal permafrost landforms such as palsas were excluded (Sollid and Sørbel, 1998), although they exist within both landmasses. Relict ice-wedge polygons are observed in northern Norway (Svensson, 1962, 1992), while no currently active examples are known to exist in either countries.

The landforms were identified based on geomorphological criteria and digitized in a GIS-environment. Each landform was characterized as either being a talus- or moraine-derived rock glacier, or an ice-cored moraine, and further attributed with characteristics on the state of activity and shape.

From the air, rock glaciers resemble both glaciers and lava streams in appearance (Barsch, 1996, Haeberli et al., 2006), and characteristically have steep front and side slopes and clear creep features at their surface such as furrows and ridges parallel to the outer landform margin. Active and inactive rock glaciers are additionally characterized by little or no vegetation, and front slopes at the angle of repose or steeper during winter (Barsch, 1996). Active rock glaciers currently creep downslope as a response to gravity, while intact rock glaciers have stagnated either due to climatic changes or topography. Active and inactive rock glaciers were here termed intact rock glaciers because of unclear borders between the two types strictly based on image interpretation. Relict rock glaciers, however, have stopped moving, and are often covered by extensive vegetation. Sometimes, but not always, taliks and collapse structures are visible at the surface, and front slopes have been worn down and are no longer standing at the angle of repose.

In light of the on-going nomenclature debate considering rock glaciers introduced in section 2.3, it is not a trivial exercise to classify the different kinds of rock glaciers in terms of origin. However, clear extinctions were obvious from the aerial photos and also acknowledged in literature (e.g. Humlum, 1982, Frauenfelder et al., 2003, Berthling, 2011), and it was ultimately decided to consider two types of rock glaciers; talus-derived rock glaciers and moraine-derived rock glaciers. Rock glaciers which origin in and creep out of talus slopes were ascribed as talus-derived rock glaciers (Figure 8a). These commonly consist of deforming pore ice which in combination with gravity is the main cause of creep. Rock glaciers which creep out of moraine ridges were termed moraine-derived rock glaciers and thus commonly exist lower in the terrain than the glaciers and regional glaciation limit, which is in accordance to Haeberli (1985). These rock glaciers might possess a deforming glacial ice-core, but the movement of the landforms are clearly decoupled from the glacier situated above (Figure 8b). Whether or not these rock glaciers consist of an inner ice-core or is cemented by pore ice is virtually impossible to decide from aerial images. For interior classification ground penetrating radar or borehole analyses is necessary, and this was not performed here.

Ice-cored moraines are formed where polythermal or cold glaciers terminate in permafrost environments. They are characterized as clearly standing out of their surroundings, and are commonly also taller than the glacier surface behind (Figure 8c). This latter characteristic is due to the general on-going glacier volume loss (UNEP, 2007). Large supplies of glacially transported material are necessary in order to build up ice-cored moraines, and they are therefore commonly found in front of cirque glaciers surrounded by weathering headwalls. This is not always the case. The material is either thrust to the surface in zones where deforming ice meets stagnating thin ice or where the geophysical properties of the ice changes from temperate to cold (Boulton, 1972). Once the material is deposited at the glacier front, it becomes part of the permafrost environment, since glacier ice is normally excluded from the permafrost definition. In the build-up phase the ice core will undergo melting during summer, but once the sediment layer is thicker than the active layer of the permafrost the ice core is protected from further melting, resulting in large moraines. Further, the sediment which belongs to the permafrost, i.e. below the active layer, is protected from evacuation from the proglacial accumulation, which additionally stabilizes the moraine compared to a non-permafrost moraine and causes additional growth. The mobile sediment of the ice-cored moraine belong to the active layer, and is affected by fluvial and colluvial processes. Ice-cored moraines were only considered to be active, i.e. intact.





*Figure 8 Examples of rock glacier types. a) Talus-derived rock glacier, ground photography taken towards south, Tröllaskagi, Iceland. b) Moraine-derived rock glacier, Storskrymten, Dovre/Sunndalsfjella, southern Norway. c) Ice-cored moraines in front of two small glaciers, Klenegga, Romsdalen, Southern Norway, aerial photos oriented towards north (© Statens Kartverk).*

Large and stable ice-cored moraines are oversized in appearance compared to moraines formed in non-permafrost environments. They possess a stable core of glacial ice, and can be recognized on air photos as disproportionately large compared to the glacier which produced them. Sometimes, but not always, ice-cored moraines show surface creep structures.

### *3.1.2 Synthetic Aperture Radar (SAR)*

Interferometric SAR or InSAR approaches to measure rock glacier velocities have proven to be successful in the European Alps in comparison to photogrammetric, geodetic and field-based methods (Rott and Siegel, 1999, Kenyi and Kaufmann, 2000, Strozzi et al., 2004). To evaluate the rock glacier deformation, and thus validate the landform activity attributed by visual interpretation of aerial imagery, the surface velocity acquired by satellite radar interferometry was used (Strozzi et al., 2004). This approach was only used for Tröllaskagi in Iceland where a pair of ALOS PALSAR data (L-band) with 46 days temporal baseline between 16 August and 1 October 2007 was available for the purpose. The use of L-band radar data ensured a good temporal phase coherence of the observation period despite of the humid climate in the study region. The radar scenes were taken in ascending orbit with a, roughly, west-east-looking sensor. This constellation allows for determination of the east-west, i.e. line-of-sight components of displacements reliably, but the displacement components in flight direction, approximately south-north, are undetectable. No suitable PALSAR data over the study site are available in descending orbit to constrain the two-dimensional direction of displacement. The topographic contribution to the interferometric phase was removed using the above 30x30m<sup>2</sup> digital elevation model. Various projections of the original line-of-sight displacements were produced such as to the horizontal of the line-of-sight and to the direction of the steepest descent. The accuracy of the measured displacements is estimated to be a few centimetres over the 46-day period or better.

### *3.1.3 Land surface temperatures (MODIS)*

For Iceland, clear-sky land surface temperatures (LST) was obtained at a spatial resolution of 1 km<sup>2</sup> from the 'Moderate Resolution Imaging Spectroradiometer' (MODIS), and was previously successfully applied for permafrost mapping in lowland permafrost areas (Hachem et al., 2008). The gridded daily L3 LST products of MODIS Terra and Aqua (MOD11A1/MYD11A1) were used, and the average clear-sky LST for the 9-year period 2003-2011 was computed. The spatial resolution of 1 km<sup>2</sup> was not sufficient to capture the

effect of small-scale topography on LST; however, larger-scale temperature patterns in the Tröllaskagi area are well reproduced. Due to the lack of *in situ* measurements of long-wave radiation in the study area, the accuracy of the LST averages cannot be benchmarked. In a similar topographic setting on Svalbard, Westermann et al. (2011) found an agreement of better than 2 °C between *in situ* measurements and MODIS LST for the snow-free summer season, whereas the average wintertime LST was significantly cold-biased due to underrepresentation of cloudy periods with warmer temperatures in LST averages (Westermann et al., 2012). Despite of considerable uncertainty, it is concluded that remotely sensed LST can give valuable indications on the larger-scale patterns of permafrost occurrence in the study area.

#### 3.1.4 Inventory statistics

The upper boundary of the rock glaciers was constrained either as identifying a clear nick point or distinct change of the slope angle, or by subjective impressions. As digitizing the landforms in a GIS was done manually, this allowed for some considerations concerning the landform constrains. This proved to be a problem especially concerning the relict landforms, where the whole landform is vaguer in appearance. In the GIS inbuilt ArcMap tools, such as ‘Area’, ‘Perimeter’, ‘Slope’ and ‘Aspect’, were used on the DEM (25 m x 25 m) within each landform polygon. Non-linear rules had to be applied to the aspect layer in order to obtain average values, and circular statistics as described in Mardia (1972) and Davis (2002) were applied (discussed in detail in Paper I). Further, a gridded regional temperature dataset of the normal period 1961-90 exists for Norway at a 1 km<sup>2</sup> resolution (Tveito and Fjørland, 1999). By introducing a general lapse rate of 0.0065 °C/m this dataset was downscaled to 25 m x 25 m resolution, using the ‘Zonal statistics as table’-tool. These statistics were all attributed to each individual landform, including the average MAAT and elevation. For Iceland, the additional mean landform velocity (PALSAR) and LST was attributed. All these characteristics including standard deviation are listed in Table 1, Paper I, and Table 2, Paper II.

Now, the landform polygons each contain information on state of activity (i.e. intact or relict), origin (talus- or moraine-derived rock glacier, or ice-cored moraine), in addition to the attributes listed above. To identify possible relationships and differences between groups and characteristics, first, a separation between regions was established (i.e. southern Norway,

northern Norway, and Iceland); second, the landforms in each region were divided by activity state; and third, by landform origin. To test differences between the regional groups, and within and between the activity classes for each region, statistical *t*-tests were performed for elevation, area, slope and MAAT. Test characteristics and results are listed in Table 2, Paper I and Table 3, Paper II.

### 3.1.5 *Geomorphic distribution models*

Analyses of which factors that control the distribution and occurrence of landforms are easier performed once a landform inventory exists. One approach is to use statistically-based geomorphic distribution models (GDMs), where field data are related to geomorphic features using empirical models. In such a statistical approach the landforms are the response variables (Hjort et al., 2007), and the goal is to provide simple relationships to explain complex relationship between process and environment. In the analyses the simplest possible relationship, using a reduced number of explanatory variables, is searched for using explanatory models where a reductionist approach is an intrinsic property. Environmental factors serve as potential explanatory variables or predictors. Further, this sort of experiment can provide tools to map previously unmapped regions for landform occurrence (Luoto and Hjort, 2005).

For testing of the GDM's ability to predict occurrence of rock glaciers and ice-cored moraines in Norway, a test area within the Lyngen/Kåfjord (Figure 9) region was selected. In this area the landform density is high, and in a grid of 500x500 m<sup>2</sup> all cells were assigned with binary values where 1 indicates landform presence and 0 that landforms are absent. Since the climate data applied as explanatory variables are modern, only intact landforms were selected.

The explanatory variables derived from the terrain model were mean elevation, slope (first derivative of elevation), aspect, curvature (second derivative of elevation), concavity, solar insolation, and elevation-relief ratio among others. Additional available environmental explanatory variables in this case were bedrock (Sigmond, 2002, Olesen et al., 2010), surface cover (Thoresen, 1991), vegetation type (Heggem and Strand, 2010), MAAT (Tveito and Førland, 1999), mean annual precipitation (MAP) (Mohr and Tveito, 2008), maximum snow depth (Engeset et al., 2004a, Engeset et al., 2004b), topography roughness index and wetness index, the latter two referred to as compound parameters (Etzelmüller et al., 2001b). The geomorphic system was then analysed by applying a general linear model, providing the most



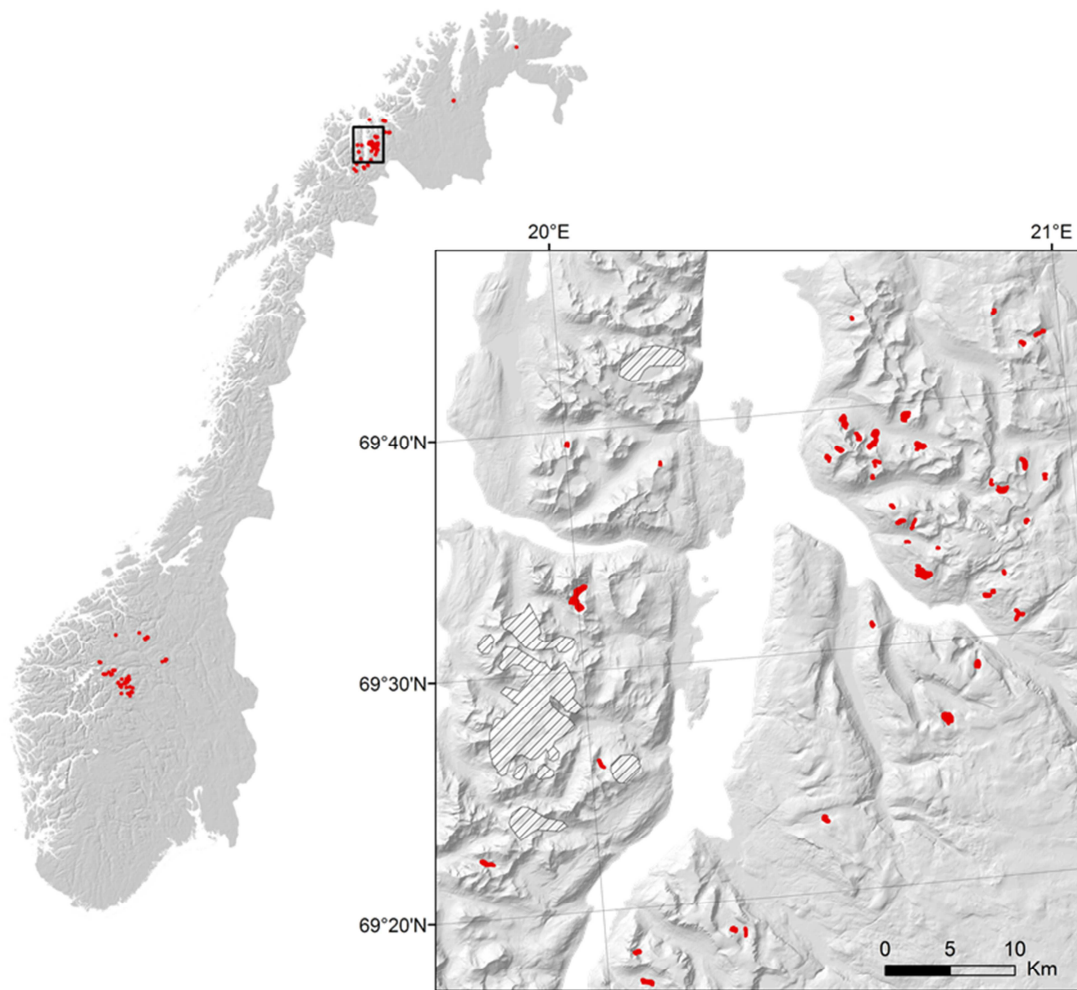


Figure 9 The Kåffjord/Lyngen area selected as a test region for a geomorphic distribution model. Intact landforms are shown in red.

important variables for landform presence to be identified. These type of models can provide insights to important factors concerning landform development, and perhaps move research efforts in more efficient directions. Modelling of presence of geomorphic landforms on a multi-variate basis may also provide higher objectivity in landform interpretations (Ayalew and Yamagishi, 2005).

The statistical software *maxent* (Phillips et al., 2006a) was used for the exercise, which was performed by J. Hjort, Professor at the University of Oulu, Finland. This software can be used for several data types and purposes, and is designed to make predictions or inferences from incomplete data (Phillips et al., 2006a). The model needs calibration, therefore 70 % of the grid cells prepared for testing was used for calibration and 30 % for

evaluation, selected at random. Where the data was incomplete (relevant for MAP and snow depth), all observations were deleted. In order to obtain more reliable results only parts of the potential explanatory variables were selected due to high intercorrelation between variables such as for example MAAT and elevation. Only linear, quadratic and terms describing interaction were fit to the curve, and not complex thresholds. The variables used in the test sample were bedrock, roughness index, MAAT, concavity, slope, maximum snow depth, solar insolation and curvature.

## **3.2 Holocene temporal and spatial permafrost modelling**

### *3.2.1 Holocene temperature series*

Little is known about the Holocene permafrost dynamics of Norway, since the permafrost has not been directly dated. However, in Norway, a network of deep and shallow boreholes (Figure 10) now exist and has been temperature-monitored for a period of between three and fourteen years, as part of the European project on permafrost monitoring in a north-south transect through Europe (PACE) and the joint Norwegian project on links between ground and air temperatures (CryoLink). A substantial amount of modelling and verification have been put into these projects, with good and recently improved 1D transient heat flow models and enhanced understanding of the ground thermal regime as main outputs. These achievements were used to evaluate the ground thermal regime at the borehole sites over the Holocene both to address the possible formation period of the currently relict and intact permafrost landforms inventoried in Paper I and to identify periods of degrading and aggrading permafrost at each borehole.

The 1D heat flow model is driven by ground surface temperatures for each year, and in order to run the model over the entire Holocene, a temperature series ultimately providing yearly average Holocene temperature data had to be compiled. Thus, available temperature records which covered the period of interest, i.e. 10,000 years BP until present were collected, manually read and plotted in 250 year time steps. The pre-existing records were organized by southern and northern Norway.

In addition, seasonal temperatures, represented by July and January temperature anomalies, were desired to address seasonal variations. Hence, mean annual, mean July, and mean January temperature anomaly series were compiled for southern and northern Norway.

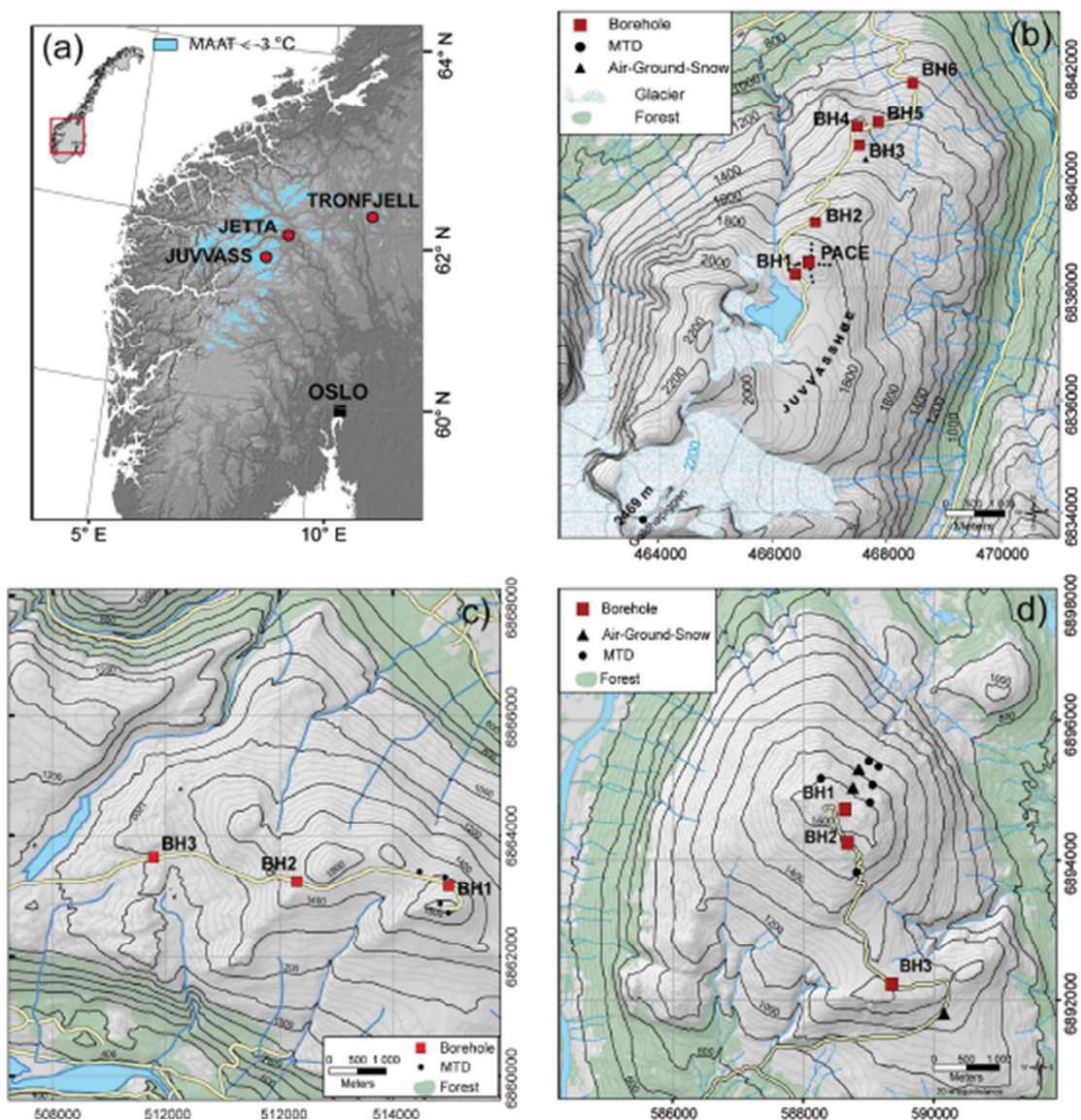


Figure 10 Borehole locations and transects in Southern Norway (CryoLink; Hipp et al., 2012).

Several proxy temperature series exist, derived from pollen and plant macrofossils in lake sediment cores (e.g. Rosén et al., 2001, Seppä and Birks, 2001, 2002, Davis et al., 2003, Bjune et al., 2004, Nesje et al., 2005, Jensen and Vorren, 2008), chironomids in sediment cores (Rosén et al., 2001, Seppä and Birks, 2001, Korhola et al., 2002), tree line changes (Dahl and Nesje, 1996, Vorren et al., 1999, Seppä et al., 2002), and deposition rate of speleothems (Lauritzen, 1996, Lauritzen and Lundberg, 1999). The use of organic material

normally retrieved via lake sediment cores has the advantage of providing datable material as well as a temperature proxy, however, not as a yearly averages but is normally associated with the temperature of the warmest month of the year, i.e. July. All July temperatures were compared using multiple regression and all records with mutual  $r^2$ -values higher than 0.7 were averaged and further considered representative of the Holocene July temperature anomaly for one region.

The resulting temperature series vary considerably between different sample sites and also between results from the same lake using different methods (Seppä and Birks, 2001, 2002, Nesje et al., 2005). Several factors can influence these results, including year-to-year variations in pollen production, in pollen dispersal and sedimentation rates, and statistical uncertainties in percentage pollen counts (Nesje et al., 2005). Vegetation also needs up to a few hundred years to adjust to abrupt climate changes, while insects are more mobile and sensitive to short-term changes, and different methods often need to be combined in order to obtain comprehensive results (Bradley, 1999). Further, each method is associated with characteristic sensitivity and response times to variations in climate. Additional uncertainties are introduced when the site in question is situated in areas close to the present tree line, where relevant pollen sources are affected by fluctuations in the timberline. For the present purpose, only major long-term trends or ‘signals’ are necessary, and large anomalies has been drawn from the different datasets and tentatively put together to one dataset.

The best mean annual temperature dataset of Norway is made from interpretation and dating of speleothem growth in Mo i Rana, Nordland county (Lauritzen, 1996, Lauritzen and Lundberg, 1999), situated in northern Norway, however geographically midway between the areas of interest in northern and southern Norway. The overall relationship with other well-known Holocene temperatures such as those retrieved from e.g. Greenland ice core data is reasonably good, but with larger amplitudes in Norway ( $r^2 = 0.55$ ) (Alley et al., 1995, O'Brien et al., 1995). The speleothem data was extended to cover the whole Holocene by use of the relationship with the GISP2 temperatures, and regionally adjusted using the July temperature dataset constructed as described above. There are some known differences in the climatic history of southern and northern Norway, first, the temperature amplitude tends to attenuate closer to the Arctic, and second, the timing of the onset of Neoglaciation is delayed in northern Norway compared to southern Norway (Bakke et al., 2005, Bakke et al., 2008).

The January temperature anomaly was calculated from linear regression of modern temperature relationships. Normally, January temperatures follow the mean annual temperature of a meteorological station much closer than the July temperature. Using data

from several meteorological stations, the best regression coefficient ( $r^2 = 0.51$ ,  $N = 250$ ) describing modern January mean temperatures of southern Norway was obtained by simple regression of the MAAT ( $T_{Jan} = a + bT_{MAAT}$ ). For northern Norway, the best ( $r^2 = 0.50$ ,  $N = 150$ ) was obtained when also including the mean July temperature in the equation ( $T_{Jan} = a + bT_{MAAT} + cT_{Jul}$ ).

### 3.2.2 Heat flow modelling

This 1D heat flow model was first implemented and used by Farbroten et al. (2007b) and Etzelmüller et al. (2011) for Iceland and Svalbard, respectively. In the model, heat conduction is assumed to be the only process of energy transfer and the heat flux equation is solved (Williams and Smith, 1989):

$$\rho c \frac{\delta T}{\delta t} = - \frac{\delta}{\Delta z} \left( k \frac{\delta T}{\delta z} \right) \quad (1)$$

where the temperature evolution of the ground (T) over time (t) and depth (z) is described. The main ground thermal properties are density ( $\rho$ ), heat capacity (c) and thermal conductivity (k). In the model, the borehole stratigraphy is implemented by using different values of the ground thermal properties for each layer. To consider changes of the latent heat of fusion ( $L$ ) connected to phase changes between ice and water, an apparent heat capacity was applied within a small temperature interval of  $\pm 0.1^\circ\text{C}$  around the freezing temperature (e.g. Wegmann et al., 1998):

$$c_{(T)} = c_0 + \frac{L}{T_1 - T_2} \quad (2)$$

Further, any effects of heat advection related to water flow in the active layer was neglected. The heat flow equation (Eq. 1) was discretized along the borehole depth using finite differences and subsequently solved by applying the method of lines (Schiesser, 1991).

### 3.2.3 Model initiation

The 1D heat flow model is driven by ground surface temperatures for each site, which are obtained from air temperatures via site specific  $n$ -factors representing the temperature offset between air and ground surface (Smith and Riseborough, 2002). In winter, the  $n$ -factor is affected by the thickness and duration of snow (annotated  $n_f$ ) while in summer, the  $n$ -factor is affected by the extent and kind of vegetation present (annotated  $n_v$ ). To consider year-to-year variability of snow thickness and vegetation density, a random variation of the  $n$ -factor was introduced, with a standard deviation of  $\pm 0.1$ .

At the lower end of the model domain (5000 m depth), the geothermal heat flux was used as the boundary condition. The geothermal heat flux was provided by the Norwegian Geological Survey (NGU) for each borehole site (Slagstad et al., 2009), and kept constant over the model period of 10,000 years.

An initial surface air temperature (SAT) was given at each borehole site to start the simulation, and in absence of more detailed information, steady-state profiles for an SAT of 0 °C were given. By choosing this value the model domain does not contain any initial permafrost and the obtained results thus represent minimum estimates of the permafrost extent.

Two numerical experiments were performed using these model constraints:

1. For the whole Holocene, we used the deviation from the normal period 1961-1990 obtained from the analysis of the Holocene temperature series ( $\Delta$ MAAT). For each borehole site  $\Delta$ MAAT was added to the MAAT of the normal period at these sites. The latter was derived by applying a constant lapse rate from a nearby weather station yielding annual mean air temperatures over the 10 kyr period, however, seasonal variations are not resolved. This run covers the whole time period and an estimation of ground temperatures during time.
2. For selected periods monthly temperatures are included to address active layer dynamics and the effects of seasonal differences in thermal conductivity and latent heat release depending on average annual water content. The monthly temperature at the ground surface was derived from superimposing a sinusoidal variation on the MAAT:

$$T_{month} = f(MAAT + A \sin(\frac{2\pi t}{P})) \quad (3)$$

where MAAT is the mean temperature of the 1961-1990 normal period based on meteorological information,  $f$  is a number between 0 and 1 where automatically generated random numbers around the respective  $n$ -factors (*standard deviation* = 0.1) were multiplied with the same  $n$ -factor to simulate natural variation,  $A$  is the amplitude,  $t$  is the time and  $P$  is the period. The amplitude was estimated based on a 2<sup>nd</sup> order relationship between the MAAT anomaly value and the range of January and July temperatures:

$$A = a MAATanom^2 + b MAATanom + c \quad (4)$$

where the coefficients are estimated for northern and southern Norway, respectively (Fig. 2b). In both cases the  $r^2$  exceeds 0.75. Eq. 3 is multiplied with this factor depending on  $T_{month}$  being above or below 0. An inbuilt MATLAB random function was used for this purpose.

The model calibration and validation were performed by T. Hipp during his Ph.D. work, and partly published in Hipp et al. (2012). All borehole models closely reproduced

recently measured ground temperatures, and were all started from the observed distribution of ground temperatures at the beginning of the observation period (500-1000 days). Values on thermal conductivity and bedrock density were either measured by the Norwegian Geological Survey (NGU) or found in literature (Williams and Smith, 1989, Sigmond, 2002, Olesen et al., 2010). The most important parameter used in the calibration process was the water content, which dampens the temperature signal in depth. The calibration results are good, and comparison of observed and modelled temperatures yield  $r^2$ -values above 0.9.

#### 3.2.4 *CryoGRID1.0 spatial modelling of Norway*

To investigate the spatial extent of permafrost over the Holocene, a recently implemented equilibrium model (CryoGRID1.0) was used (for details, cf. Gissnås, 2011). This model is based in the TTOP approach originally developed for Canada (Smith and Riseborough, 1996), and models the relationship between climate and permafrost. In CryoGrid1.0 three vertical layers is considered: (1) mean annual air temperature (MAAT), (2) mean annual ground surface temperature (MAGST), and (3) the temperature at the top of the permafrost (TTOP) or alternatively at the bottom of the seasonal frozen layer (MAGT) (Figure 11). The temperature gradient between the MAAT and the MAGST is normally positive, and this difference is termed the ‘surface offset’, and expresses the influence of surface cover at ground surface temperatures. The temperature difference between the upper two layers are overcome by introducing scaling factors as expressions of summer vegetation ( $n_t$ ) and winter snow cover ( $n_f$ ) termed thawing and freezing  $n$ -factors, respectively (Lunardini, 1978). If permafrost is present, the temperature gradient between the MAGST and the TTOP is negative, termed the ‘thermal offset’ ( $\Delta T_k$ ) (Goodrich, 1982, Burn and Smith, 1988, Smith and Riseborough, 1996), and is related to different heat conduction in frozen and thawed ground. Frozen ground has typically higher thermal conductivity than thawed ground. Hence, permafrost can be present where the MAGST is close to or above 0 °C.

The input parameters for CryoGRID1.0 are gridded snow and air temperature data (degree-days) provided by the Norwegian Meteorological Institute, available at 1 km<sup>2</sup>, and is validated against numerous permafrost observations in Norway (Gissnås, 2011). To address the Holocene permafrost distribution and variation of Norway, three climatically different time periods were selected; (1) the Holocene thermal maximum (HTM) (2) the LIA and (3) present. For each run the positive and negative degree-day sum was adjusted compared to the

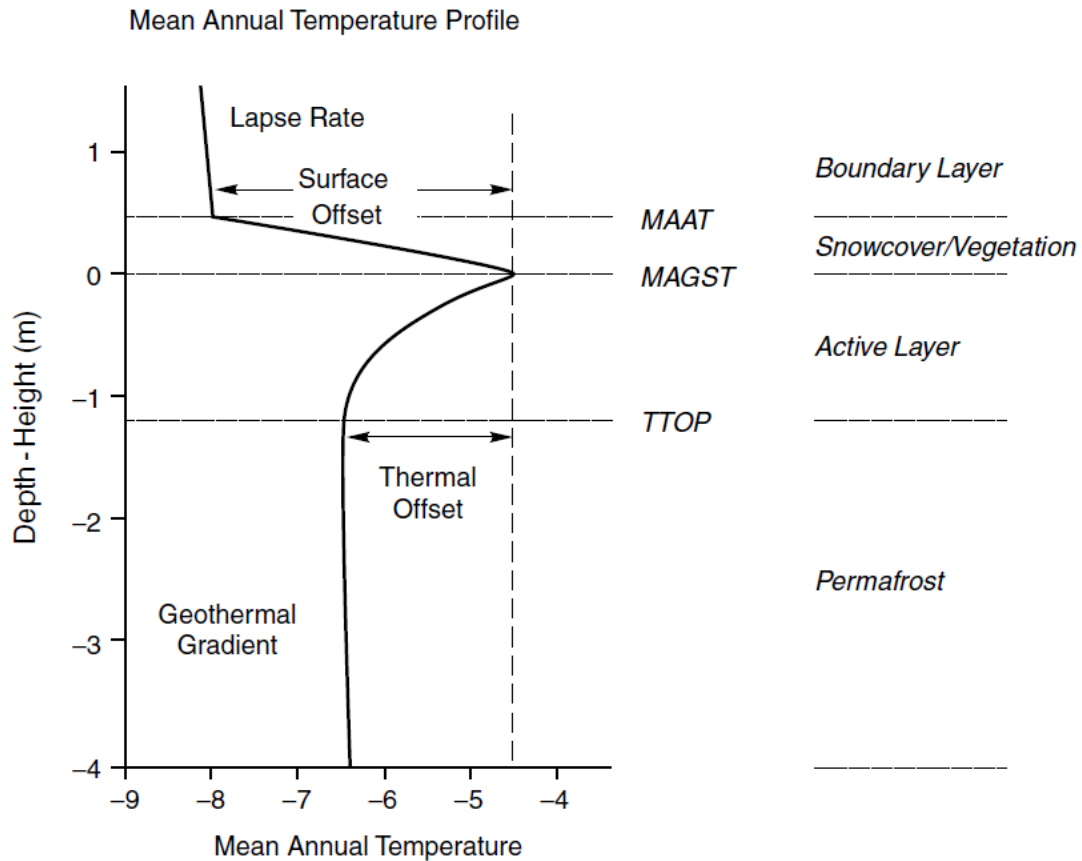


Figure 11 Schematic illustration of mean annual temperature relationships in a vertical ground profile (Smith and Riseborough, 2002). MAAT: Mean Annual Air Temperature, MAGST: Mean Annual Ground Surface Temperature, TTOP: Temperature at the top of permafrost.

present situation, and one value was used for all of southern Norway and a different value for northern Norway (Table 3, Paper III). Degree-days for the HTM and LIA are based on the relationship between modern MAAT measured at meteorological stations and the calculated freezing degree-days (DDF) from the same stations ( $r^2 > 0.85$ ), whereas the corresponding thawing degree-days (DDT) were calculated from the relation to DDF as:

$$DDT = (MAAT P) + DDF \quad (5)$$

where  $P$  is the period of one year (365 days) (Smith and Riseborough, 2002). Likewise the precipitation and hence snow depth was altered to simulate former conditions. The change of precipitation was given as a rough estimate based on e.g. Bjune et al. (2005), Matthews et al. (2005), and Nesje et al. (2001). In addition, for the LIA the vegetation was kept similar as at present, while for the HTM the timber line was raised with 200 m in altitude over the whole



country, following Dahl and Nesje (1996), Seppä and Birks (2001, 2002) and Seppä et al. (2002). In this spatial model, no 3D-effects were considered.

### 3.3 Glacial reconstruction and geomorphology, Omnsbreen

Reconstruction of the former maximum extent of Omnsbreen was done strictly by considering dynamical criteria, since the formerly glaciated zone is not constrained by landforms such as terminal moraines. However, a glacial geomorphic mapping of the area was performed, and based on landforms like striae, flutes, eskers, crag-and-tails, boulder tracks in the till, the peculiar “domino”-structures (described in Paper IV) and to a limiting degree morainic deposits an outline of the former glacier was suggested. From this starting point, surface contour lines at 25 m spacing were digitized in a GIS environment (ESRI ArcMap©), based on the assumption that the steepest surface slope is aligned parallel to the direction of the glacier movement, thus is the contour lines oriented perpendicular landform indicating ice movement direction. The glacier surface, slope and volume was calculated via inbuilt interpolation, slope and raster calculation tools (Hutchinson, 1989, Hutchinson and Dowling, 1991). Further, the subglacial shear stress ( $\tau$ ) was calculated from the equation:

$$\tau = F\rho gh \sin \alpha \quad (6)$$

where  $F$  is a shape factor assigned to valley glaciers (we used 1),  $\rho$  is the ice density (900 kg/m<sup>3</sup>),  $g$  is the gravitation (9.81 m/s<sup>2</sup>),  $h$  is the glacier thickness and  $\alpha$  is the surface slope (Nye, 1965, Paterson, 1994). A general ground resolution of 25 m was applied. The spatially distributed values of basal shear stress can be interpreted as a measure of glacier activity. Further, the strain rate ( $\dot{\epsilon}$ ) of the ice is given by Glen’s flow law:

$$\dot{\epsilon} = A \tau^n \quad (7)$$

where  $A$  is a constant which decreases at lower temperatures and  $n$  is the flow law exponent, normally with a value close to 3 (Glen, 1955, Benn and Evans, 1998). The deformation velocity ( $U_d$ ) is found by integrating the strain rate over the ice thickness, following Nye (1965):

$$U_d = \frac{2A}{n+1} (F\delta g \sin \alpha)^n h^{n+1} \quad (8)$$

Considerations concerning sliding velocities were not applied.

## Chapter 4

# Results – Summary of papers

### 4.1 Inventories of permafrost landforms – Norway and Iceland

#### 4.1.1 Norway (without Svalbard) (Paper I)

In total, 307 permafrost landforms were identified in Norway, distributed by region, activity state and genesis as described in Table 1 and Fig. 1, Paper I. One of the main outcomes of this inventory is the observation that most relict landforms are of a talus-derived type while the majority of the currently intact landforms are connected to modern glaciers, either as ice-cored moraines or as moraine-derived rock glaciers. From this observation it is interpreted that a process-shift concerning the formation of these landforms occurred at some point during the Holocene, where the modern landforms are mainly connected to glacial activity. This observation does not contradict the definition of rock glaciers as creeping permafrost. The most likely timing of this process-shift is associated with the Neoglaciation, when the climate after a warm and dry period turned colder and wetter. This climate change led to the reappearance of the Scandinavian glaciers after the deglaciation, thus ‘Neoglaciation’. The glacier-related landforms also represent the majority of the total amount of inventoried landforms, regardless of activity state (Figure 12).

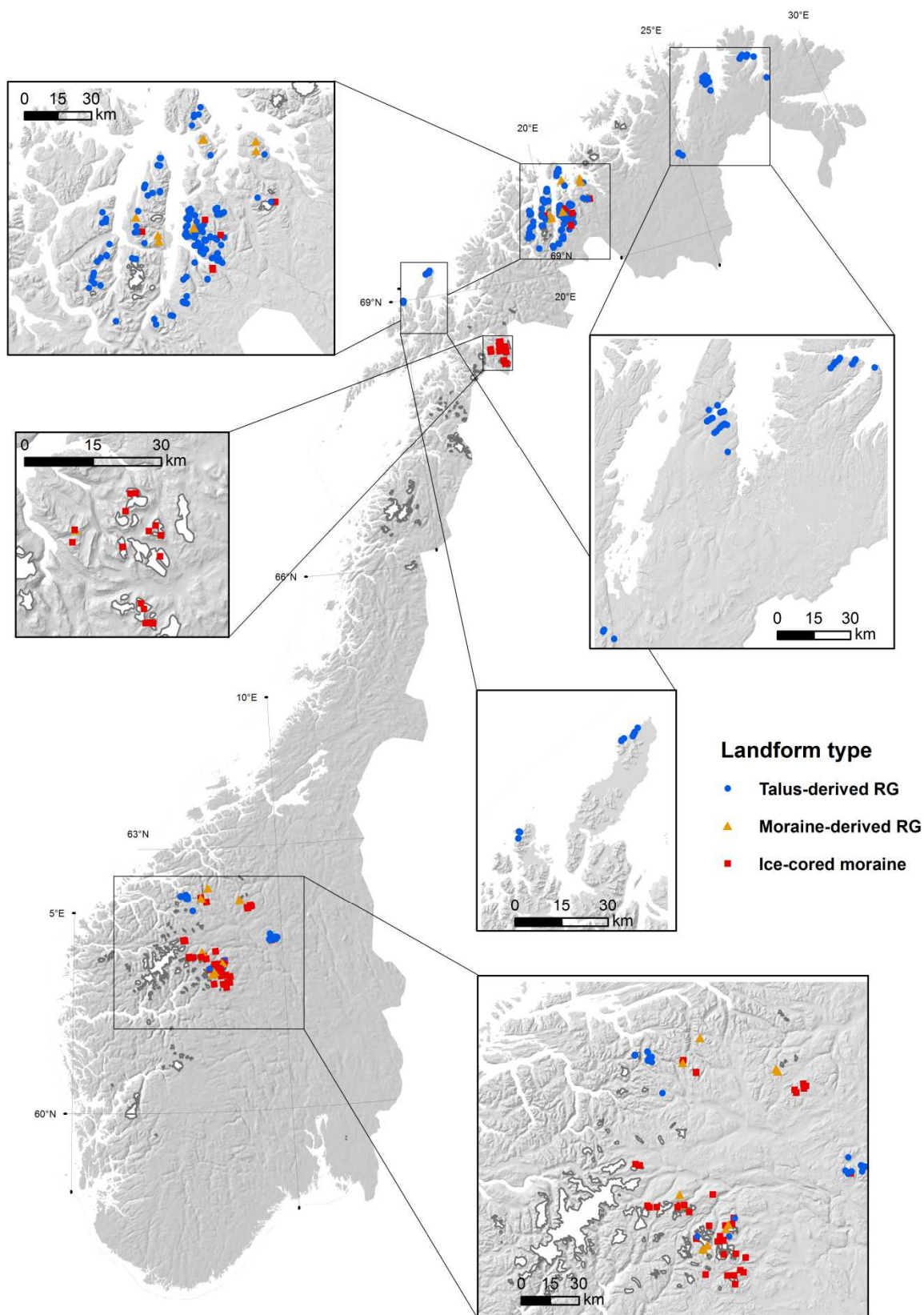


Figure 12 Inventoried permafrost landforms of Norway, classified by landform type (genesis).

Of the inventoried landforms, the intact landforms are strongly dependent on aspect, while this is not the case for the relict landforms. This is a logic implication of former permafrost being a widespread ground thermal phenomenon along the ice-sheet margins during the deglaciation, whereas modern permafrost is a relatively marginal high-mountain phenomenon in comparison and depend on altitude, solar radiation and specific surface heat fluxes to exist. This is also why the relict landforms are situated at all elevations, including at sea level.

The permafrost landforms inventoried were treated as different landform populations. The physical distance between the landform clusters of southern and northern Norway is long, and differences in statistical characteristics such as mean elevation and temperature would otherwise alter the result. There are also additional differences between the populations, for example the vast majority of the landforms in northern Norway are relict, while the opposite is the case in southern Norway. In northern Norway, the most frequently occurring intact landforms are derived from talus slopes, while in southern Norway ice-cored moraines occur most often. One suggested explanation for this observation would be that modern permafrost landform formation in maritime permafrost environments is strongly connected to glacial activity, manifested as moraine-derived rock glaciers and ice-cored moraines instead of 'dry' talus-derived landforms. The same observation is done in Iceland, cf. section 4.1.2.

In literature, the activity state of rock glaciers has been used to validate mountain permafrost zonation, e.g. active rock glaciers belong to areas of at least discontinuous permafrost and relict rock glaciers to areas of former permafrost (e.g. Imhof, 1996, Lambiel and Reynard, 2001, Janke, 2005). As validation for existing Norwegian permafrost models (e.g. Ødegård et al., 1996, Etzelmüller et al., 2001a) the inventory only works to a limited degree. In Norway, the number of rock glaciers is low compared to regions like northern Iceland (Paper II), the European Alps (Frauenfelder et al., 2003), the Colorado Front Range (Janke, 2007), or the Andes (Brenning, 2005), and a statistical approach to validate permafrost models would not give significant results on landscape or mountain range scales. However, the lower inventoried landforms and the internal zonation between activity groups is within the permafrost models' predictions. In addition, the distribution of permafrost is affected by 3D-effects, and will penetrate both deeper in the ground and lower in the terrain in north-facing compared to south-facing slopes in the northern hemisphere (Noetzli and

Gruber, 2009). As the modern rock glaciers strongly depend on aspect, the lowest situated examples represent an absolute lower limit of modern Norwegian permafrost.

4.1.1 b) *Applications of a geomorphic distribution model (not included in paper)*

For further and perhaps more objective classifications of the decisive environmental parameters causing rock glaciers to form, a test run using the statistical software *maxent* was applied on a subsample of the Norwegian intact rock glacier population. Only intact rock glaciers were considered since the climatic parameters applied are modern. The parameter which proved to be by far the most important predictor for rock glacier presence within the grid cells was bedrock, followed by the landscape roughness index and MAAT (Table 1). The landscape roughness index expresses the topographic variation in an area, and influences the amount of incoming radiation (Etzelmüller et al., 2001b). Bedrock as an explanatory variable also possesses the highest importance when used in isolation, and has the highest impact on the result when omitted from the model (Figure 13). Bedrock is therefore considered to hold the most information that is not present in other variables.

*Table 1 Contribution and permutation importance of the environmental variables considered for prediction of rock glacier occurrence.*

<b>Variable</b>	<b>Percent contribution</b>	<b>Permutation importance</b>
Bedrock	41.5	48.5
Roughness	17.2	8.2
MAAT	14.3	25.2
Concavity	14.2	9.6
Slope	9.2	2.2
MaxSnowDepth	1.5	5.2
MAP	1.1	0
SolarRadiation	0.8	0.8
Curvature	0.2	0.4

Bedrock was not considered as a variable in the statistical analyses performed in Paper I. In this paper only MAAT and the topographic variables elevation, slope and aspect were used to reveal differences between landform activity, origin and populations. There was a high internal dependence on aspect for the intact rock glaciers, whereas in the GDM aspect is

‘translated’ into solar radiation, which has the second lowest predictive power concerning presence or absence of rock glaciers in Lyngen. Thus, the independent validation of geomorphology interpretation that the GDMs offer can potentially be of great importance.

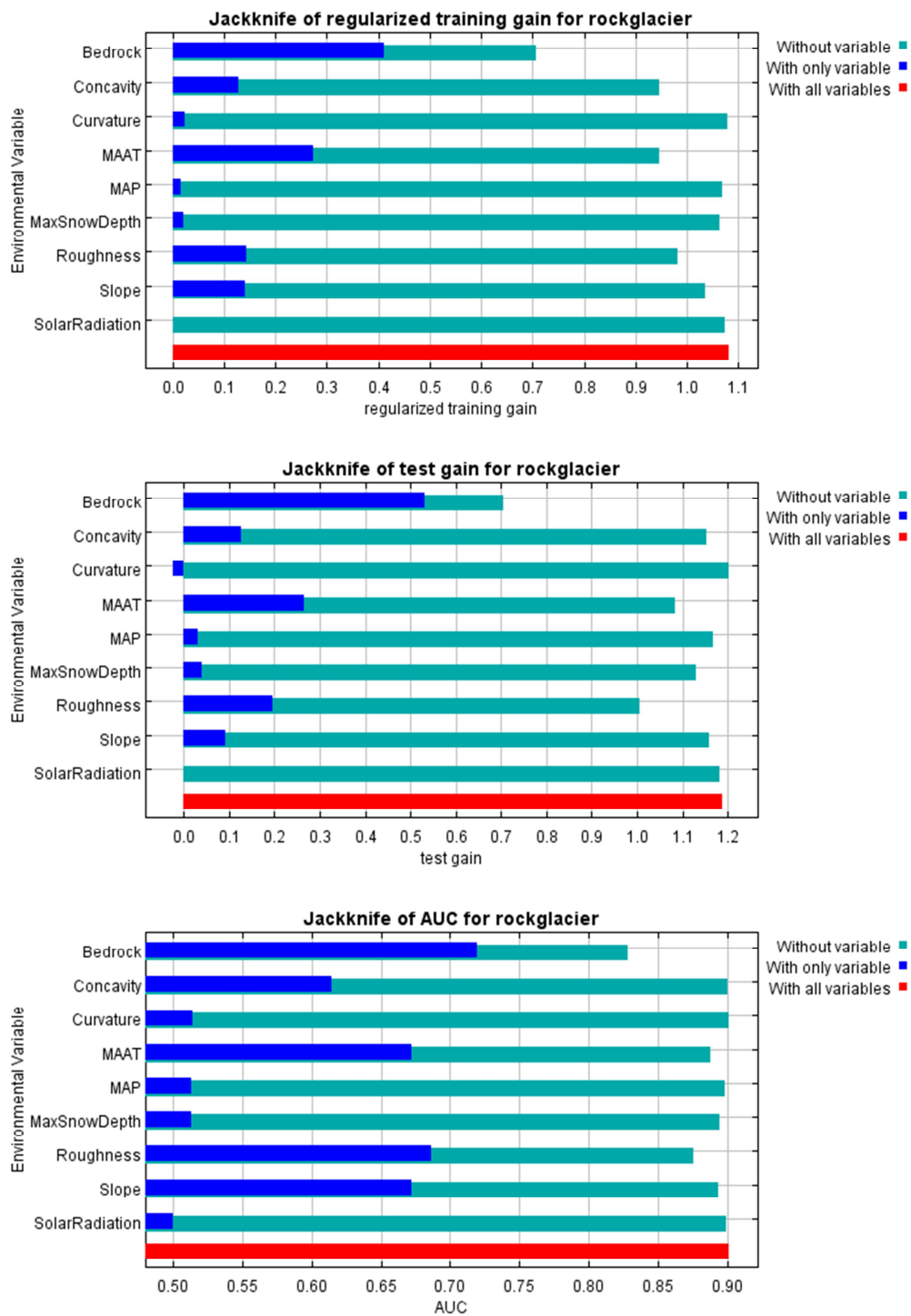


Figure 13 Jack-knife tests of the explanatory variables considered: a) for the sample used as a training sample for validation, b) for the test sample used for modelling and c) for the total area under curve (AUC). Green lines indicate the prediction power when the variable is left out of the model, blue lines indicate the prediction power when the variable is used as the only variable of the model, and red lines are the total model gain.

#### 4.1.2 *Iceland (Paper II)*

In Tröllaskagi, northern Iceland, rock glaciers and ice-cored moraines occur in abundance, and 265 landforms were inventoried within two relatively small adjacent peninsulas (refer to Table 2, Paper II for characteristics). Like in Norway, the most commonly occurring landforms are those related to glaciers, an observation which is interpreted as manifestation of maritime permafrost, and in particular the close interaction between permafrost and glacial processes (Figure 14). Similar landforms are observed in Disko, western Greenland (Humlum, 1982, 1988), but is not largely reported from for example the European Alps.

InSAR data were used for additional information on movement and rates of movement in the inventoried dataset. The characterization based on image interpretation coincided with the produced landform velocities, and was therefore not altered. Actually, the results from InSAR data served as verification of the image interpretation approach initially used to classify landforms in terms of activity both in Iceland and mainland Norway. The landforms velocities received were up to  $1.2 \text{ m a}^{-1}$  (Figure 3b, Paper II), and in good agreement with measurements on individual landforms by GPS and photogrammetry (Whalley et al., 1995b, Wangenstein et al., 2006). The land surface temperatures obtained from MODIS satellite data also support the activity interpretation of the landforms, although these temperatures appear somewhat too low in places (Figure 14). This can be both due to a bias within the timing of cloudy conditions over the year and that temperatures on top of snow covers are included in the measurements (Westermann et al., 2012). However, the overall representation is good.

Further, also in Iceland relict rock glaciers exist at low altitudes. For rock glaciers to develop into the sizes that were observed at the coast, several millennia of low temperatures are required (Barsch, 1996, Frauenfelder and Kääb, 2000). This observation supports the current interpretation of a very rapid deglaciation of Iceland. The coastal areas of Tröllaskagi were ice-free already during Bølling interstadial (c. 14,500 years BP; Ingólfsson et al., 2010) mainly as a dynamic response to the rising global sea level from a largely marine-based ice sheet. In addition, the Icelandic crust is highly sensitive to glacial loading and unloading, and compensate fast by isostatic rebound (Ingólfsson et al., 1995, Ingólfsson and Norddahl, 2001). Dry land was available for rock glacier development and subject to still cold conditions until the complete deglaciation of Iceland at approximately 8,700 years BP (Ingólfsson et al., 2010). This deglaciation was interrupted by the Younger Dryas stadial,

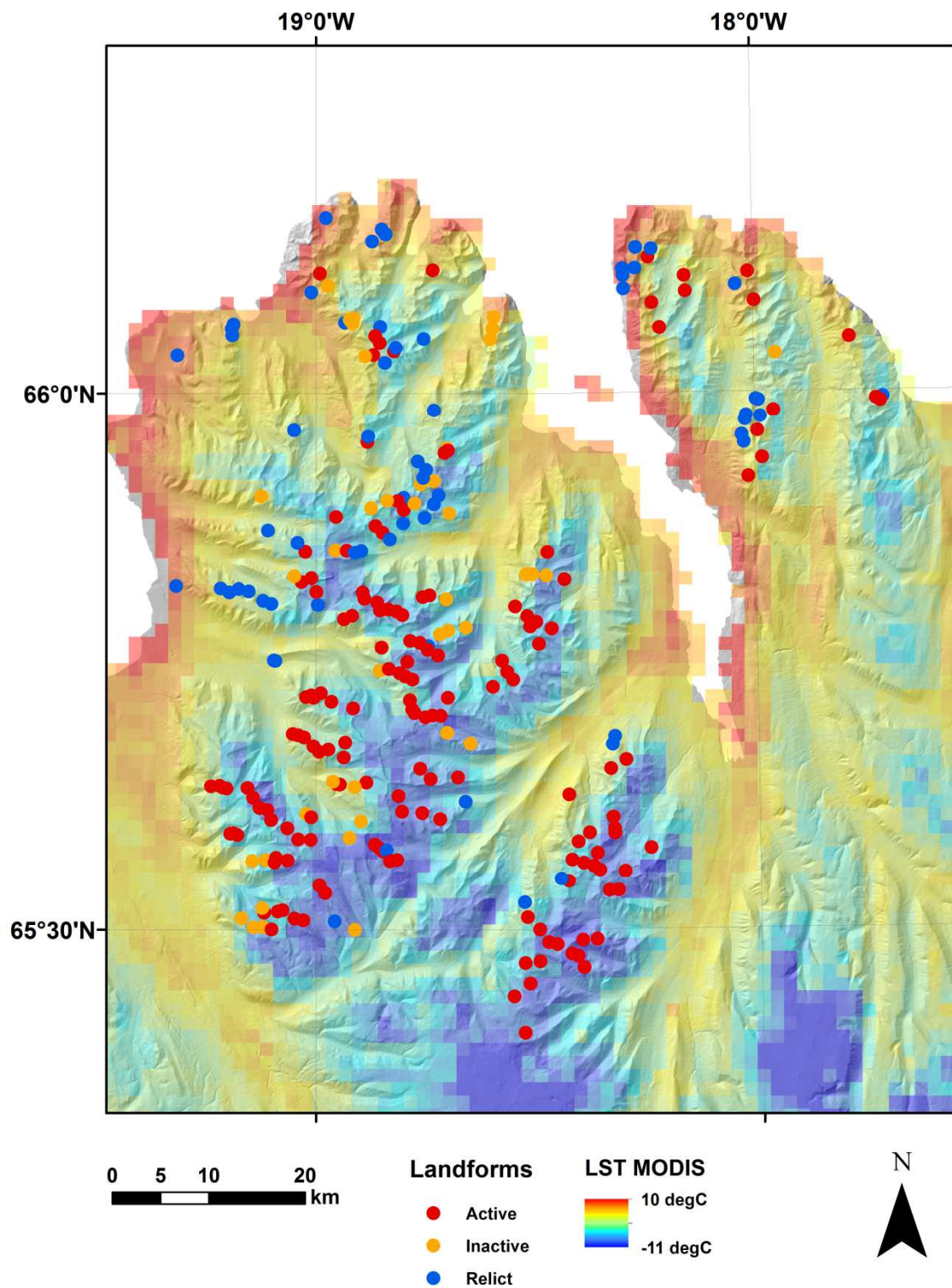


Figure 14 MODIS land surface temperatures in Tröllaskagi, with landforms and activity class indicated.



which, however, is weakly constrained in the Tröllaskagi peninsula (Norðdahl et al., 2008). In addition, the relict coastal landforms are situated at seafacing slopes and not fjordfacing slopes where they probably would be affected by advancing glaciers. In fact, the lack of relict landforms in such slopes might suggest possible removal of landforms by glaciers. Two places in Tröllaskagi rows of relict moraine-derived rock glaciers in cirques at 4-500 m a.s.l. are observed, which is too low for current permafrost. However, one could speculate that these rock glaciers were active during the Younger Dryas and as such represent reglaciation of low-elevation cirques, whereas on the other hand an LIA age is also plausible.

The currently active landforms in Iceland are interpreted to be of mid-Holocene age and Neoglacial origin (Wangensteen et al., 2006, Farbrot et al., 2007a, Kellerer-Pirklbauer et al., 2008). Also, the rock glaciers display relatively high surface velocities, which indicate 'warm' permafrost and/or high ice content (Kääb et al., 2007). Active rock glaciers are the cumulative expression of their entire life span, and develop through creep over millennia. In addition, the age is often considered proportional to the rock glacier size. For relict rock glaciers, a late-glacial age is often inferred (Barsch, 1996, Frauenfelder and Kääb, 2000). Obviously, rock glacier initiation requires ground permafrost conditions. In addition sufficient material supply is necessary, which also serve as a delimiting factor in some regions. The latter, however, is not the case in Iceland.

## **4.2 Relative age of Holocene permafrost in Norway (Paper III)**

The major findings when modelling Holocene permafrost in mainland Norway are the permafrost persistence during the HTM in high altitudes, and that the greatest extent of permafrost both in distribution and in depth is found during the LIA. From these observations, five altitudinal zones of relative permafrost age is suggested; (1) in the highest altitudes the permafrost predates Holocene, and at subsequently lower altitudes (2) taliks formed above permafrost during the HTM, (3) permafrost thawed during HTM, and permafrost postdate the HTM, (4) permafrost aggraded during the LIA, while is currently under degradation and (5) permafrost was never present over the Holocene. Figure 15 shows the TTOP/MAGT at six different times during the Holocene, for a small area in southern Norway.

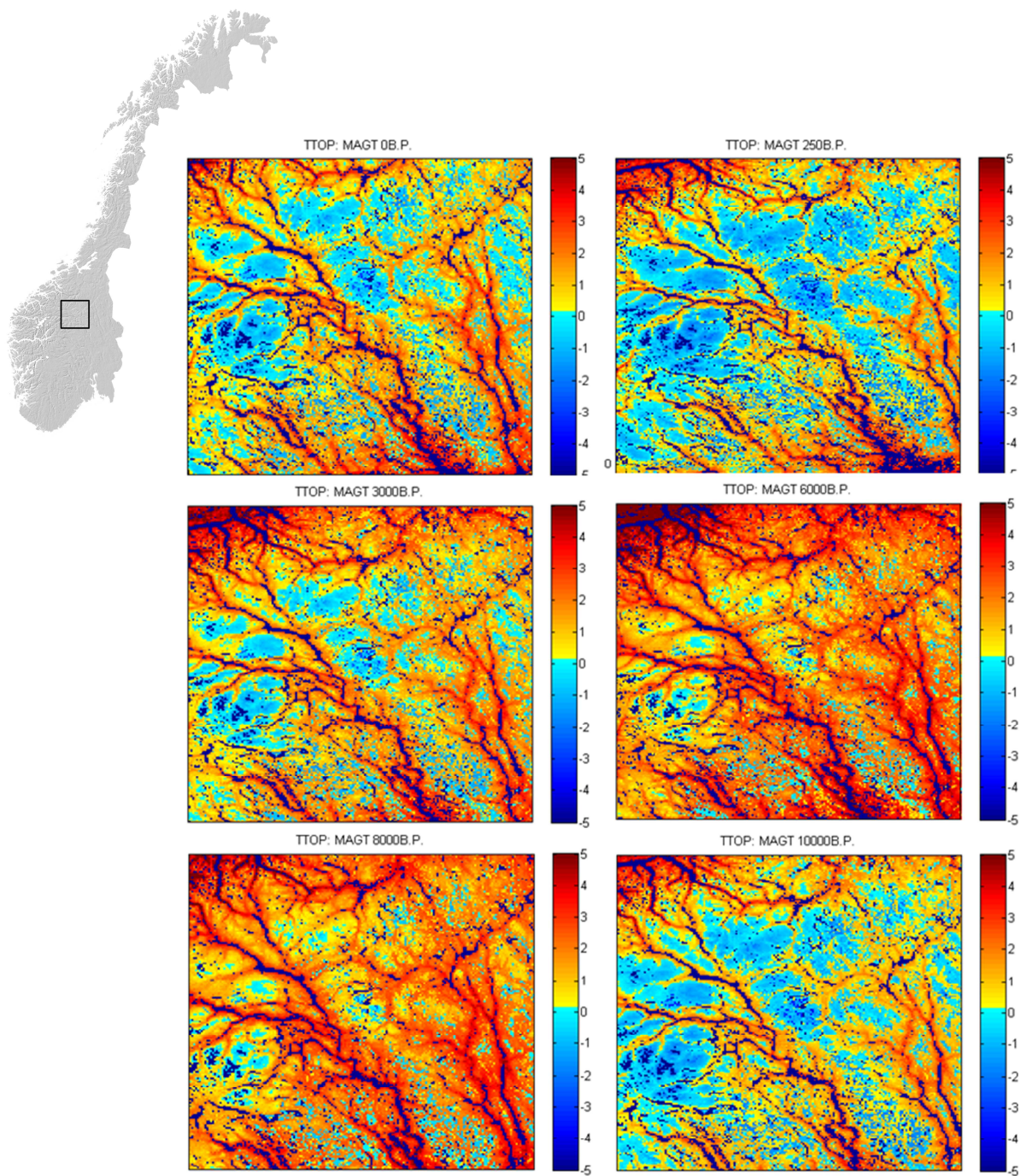


Figure 15 Mean annual temperature at the top of permafrost (TTOP) and bottom of the seasonally frozen layer (MAGT) as modelled by CryoGRID1.0 at six different times during the Holocene. The upper left corner is the present (1981-2010) situation, and the lower right corner indicate the situation by the end of the last glaciation. Dark blue colours indicate permafrost, whereas yellow and red colours are non-permanent frozen ground.

When comparing the modelled permafrost distribution during LIA with the current distribution, the highest sensitivity is found in northern Norway, and in particular the northernmost county, Finnmark. Here, the permafrost distribution is now reduced by more than 50 %, and the area might represent a link between the mountain permafrost of Scandinavia and the continuous Arctic permafrost. Most of the areal reduction, however, is found at Finnmarksvidda, a large low-relief plateau apparently situated at an elevation which favoured permafrost during the LIA climate, while it no longer does. The significance this represents in terms of permafrost sensitivity is large.

During the Quaternary glaciations, large areas were underlain by permafrost in Fennoscandia. Areas that traditionally has been ascribed as covered by a cold-based ice sheet are areas associated with landforms such as lateral meltwater channels, Rogen (ribbed) moraines and extensive sediment cover (Sollid and Sørbel, 1984, Kleman, 1992, 1994, Kleman and Borgström, 1994, Sollid and Sørbel, 1994, Hättestrand, 1997, Kleman and Hättestrand, 1999). The distribution of these areas compared to the LIA distribution of permafrost as modelled by the CryoGRID1.0 model is striking. However, the CryoGRID1.0 approach is more detailed, and restricts these areas to a larger degree than the traditional zones of areas of cold-based ice, but is of course greatly underestimated compared to the situation at the onset and during the last glaciations. Often is the occurrence of cold-based ice associated to the deglaciation and a thin ice-sheet, due to the often co-existing or peripheral existence of landforms indicative of a temperate and eroding ice-sheet.

### **4.3 Glacier-permafrost interactions, exemplified by Omnsbreen (Paper IV)**

As shown in the previous section based on Paper III, permafrost was likely to occur under a much larger area during the LIA than at present. Simultaneously the glaciers gained mass and aggraded into new areas, creating a large potential of permafrost and glaciers to coexist and respectively related processes to interact in Norwegian mountains. Areas peripheral to the current glaciers and permafrost areas were affected by glacier-permafrost interactions in that period. In this respect, the currently small (<0.5 km<sup>2</sup>), but previously much larger, glacier Omnsbreen, situated north of Finse in southern Norway represent a kind of glacier which both formed and almost completely disappeared during the LIA. Today, the glacieret is

located in the western slope of a north-south trending valley, and is completely dependent on accumulation of wind-redistributed snow to exist. The glacier's equilibrium line altitude (ELA) is currently situated below the regional ELA at c. 1700 m a.s.l. At its maximum extent, the glacier was by glacial-geomorphic criteria mapped to have covered an area of 7 km<sup>2</sup>. The first appearance of Omnsbreen in the catchment of the nearby Lake Omnsvatn occurred at AD 1425 ± 85 years (Sægrov, 2006) which is in close agreement with radiocarbon datings (1430 ± 100 years) of plant remains covered during the formation of Omnsbreen (Elven, 1978). These dates correspond to the onset of the LIA. In early-LIA a shift in wind patterns occurred, and traces of more frequent winter storms are found in Greenlandic ice cores (Kreutz et al., 1997, Fischer et al., 1998). The current glacieret is entirely dependent on wind-redistributed snow to exist, and a change in wind fields might have given the lee areas an extra amount of winter snow during the LIA. Based on observations on radial striation, and current and previous wind patterns of the area, it is suggested that the glacier was initiated at the same spot where it is located today, and subsequently grew from this spot to its maximum size centuries later. At its maximum size, the glacier presumably filled the whole valley of Omnsbreen, and could to much less degree than previously benefit from its lee-side occupation. Rather, the glacier might have been subjected to wind erosion at its surface at this maximum stage. The glacier had a fast but late retreat, with the main mass loss occurring between 1930 and 1985, until it reached its approximate current size. Again, the glacier was positioned in a lee-side slope.

The CryoGRID1.0 model predicts sporadic permafrost in the glacial area at present and widespread permafrost during the LIA. Contemporary permafrost is observed as frozen subglacial till below the nearby outlet glacier Midtdalsbreen, a northwards draining outlet glacier of the much larger Hardangerjøkulen ice cap, and by DC-resistivity measurements (Liestøl and Sollid, 1980, Etzelmüller et al., 1998). In the late summer of 2011 an attempt to replace a malfunctioning ground temperature logger at 1550 m a.s.l. was unsuccessful due to frozen ground at 30 cm depth. The winter of 2010/2011 was less snow-rich than previous years and illustrates the sensitive ground thermal regime of the area, and the high dependence of snow distribution. Based on these observations, it is probable that during the formation of Omnsbreen in the 15<sup>th</sup> century permafrost had formed in the ground, and Omnsbreen aggraded into a permafrost environment. As the glacier grew, the basal ice turned temperate, and the glacier remained a mainly temperate glacier throughout its life span. Marginal zones, however, were cold.

## Chapter 5

# Overall discussion

### 5.1 Permafrost landforms – Norway and Iceland

The most important findings considering the permafrost landform inventories are first the very low abundance of permafrost landforms in Norway and the very high abundance in Iceland compared to the European Alps, western Greenland, the Colorado front range etc., and second the dissimilarity concerning landform origin between the early-Holocene landforms and the mid- to late-Holocene landforms.

#### 5.1.1 *Permafrost landform abundance*

Despite the rather widespread occurrence of permafrost in mountainous Norway, especially active talus-derived rock glaciers are almost non-existent. This situation can be a result of high-competent bedrock not prone to weathering, leading to a debris-limited distribution and occurrence pattern of landforms. This assumption is also supported by results of the GDM performed by J. Hjort, where bedrock turned out to possess the highest explanatory power for rock glacier occurrence. In Iceland, the bedrock is young, highly prone to weathering and situated within an active tectonic environment causing large amounts of debris available for deformation when subjected to permafrost. There, the geomorphic expression of permafrost is not limited by the availability of debris, and the patterns of activity zones are more likely to realistically represent the permafrost distribution. This ‘Icelandic situation’ is also the case elsewhere where the activity of landforms have been used to reflect the zonation within

mountain permafrost, such as in the tectonically active European Alps (Frauenfelder et al., 2001), the South American Andes (Brenning, 2005), and perhaps also in the less studied, in terms of permafrost geomorphology, south island of New Zealand.

### *5.1.2 Change of permafrost landform origin during the Holocene*

The relict permafrost landforms exist at all altitudes, and are primarily formed from talus slopes or as secondary creep in landslides invoked by isostasy-driven earthquakes. These landforms thus reflect a severe climate, and dry, periglacial environments. Both the inventories of mainland Norway and Iceland suggest that the most often observed modern permafrost landforms are those which are associated to modern glacial activity, i.e. ice-cored moraines and moraine-derived rock glaciers. The few examples of moraine-derived rock glaciers which are indirectly dated in Iceland are of ages of around 5000 years (Farbrot et al., 2007a), which correspond to the general onset of Neoglaciation both in Tröllaskagi and in Scandinavia (Norðdahl et al., 2008, Nesje, 2009). In Norway, despite datings, the close relationship between glaciers and permafrost landforms is striking and a connection between the Neoglaciation and the development of glacially influenced permafrost landforms was suggested in Paper I. According to the spatial Holocene permafrost model presented in Paper III, permafrost aggraded following the HTM in mountain regions as a response to the Neoglacial cooling. Thus, the potential of glacier-permafrost interactions of the late-Holocene was large.

The permafrost regions of both Norway and Iceland must be considered maritime and warm, and the suite of landforms associated with such permafrost environments differ from the very well investigated permafrost geomorphology of the European Alps. The Scandinavian, Icelandic and western Greenlandic permafrost landform assemblage represent maritime permafrost geomorphology, where the glacier-permafrost interfingering is defining for the suite of landforms present.

Glacial influence on the early-Holocene permafrost landforms both in Norway and Iceland is much less obvious, and a change in the most important processes leading to rock glacier formation is suggested. Thus, the importance of glacier-vicinity and process interaction between glaciers and rock glaciers has traditionally been underestimated in modern maritime permafrost regions, which has further led to misinterpretations of landforms indicative of permafrost in regions like Norway and Iceland. The “ice-cored rock glacier” of Nautárdalur, Iceland, serve as an example where a correctly identified rock glacier originally and persistently was interpreted as decoupled from the permafrost regime (Whalley et al.,

1995a). Also the inventory of Scandinavian ice-cored moraines compiled by Østrem (1964) contains numerous examples of rock glaciers, as was later acknowledged by the author (Østrem, 1971). However, permafrost presence was never mentioned. In this respect it should be noted that Scandinavian permafrost and related landforms except for palsas was not recognized to be of importance until the early 1980s (King, 1983).

I suggest that a shift in the most important processes leading to the formation of rock glaciers occurred over the Holocene. In early-Holocene a dry, and periglacially dominated regime favoured formation of ‘dry’ rock glaciers, i.e. of the talus-derived type, whereas a humid, and glacially dominated regime characterising the mid- and late-Holocene favoured ‘moist’ rock glacier formation, namely of the moraine-derived type, and correspondingly to the high occurrence of present-day ice-cored moraines (Figure 16). Thus, modern and active rock glaciers in a maritime permafrost environment are likely to be moraine-derived.

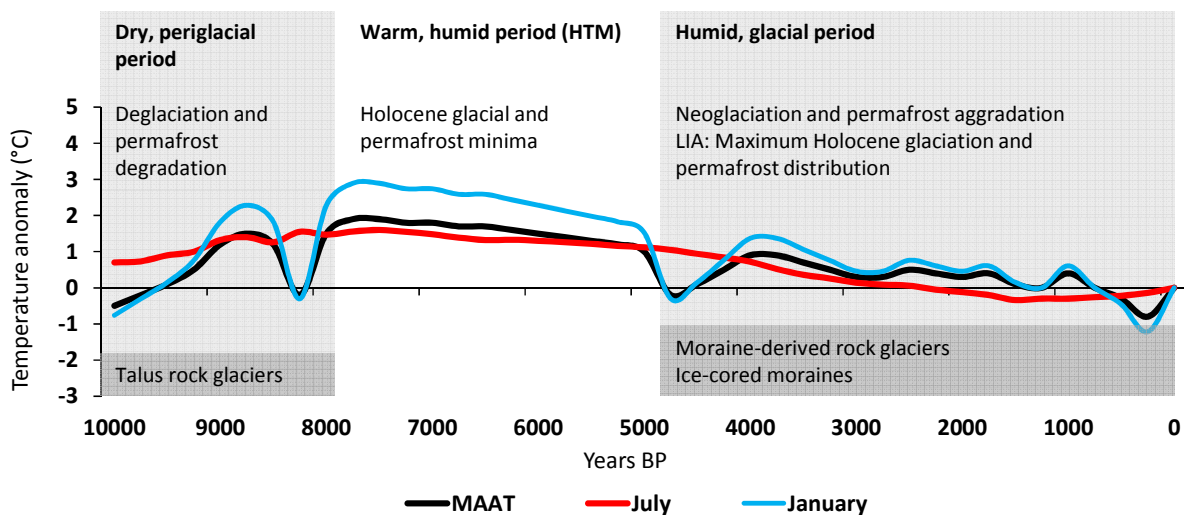


Figure 16 Holocene temperature series (Southern Norway) compiled for and applied in Papers III and IV. Distinct periods of geomorphic importance are indicated.

## 5.2 Implications for the extent of the last glacial maximum (LGM) and deglaciation of northern Iceland

The landform inventory of the Tröllaskagi region, Iceland, revealed that relict rock glaciers exist at all altitudes, including at sea level. One explanation of the formation of low-level rock glaciers is to assume a short time period, for example by secondary creep in rapidly

deposited colluvial debris. Landslides can be triggered as a paraglacial response soon after or during deglaciation of an area (Whalley et al., 1983, Dehls et al., 2000, Ballantyne, 2002). Many of the currently inactive or relict rock glaciers of northern Norway are for example interpreted to first have been deposited by landslides in permafrost environments (Tolgensbakk and Sollid, 1988), and considering the rapid isostatic rebound following the glacial unloading (Ingólfsson et al., 1995, Biessy et al., 2008), landslide triggering appear likely. The partly controversial low-land debris bodies occurring in large volumes in the Tröllaskagi regions is also commonly interpreted to reflect paraglacial stress release (Jónsson, 1976, Whalley et al., 1983).

However, in Paper II it was argued that the commonly accepted deglaciation model opened a time window of app. 6000 years since the deglaciation commenced in the Bølling interstadial (15,400 years BP) until the complete deglaciation at around 8,700 years BP (Kaldal and Víkingsson, 1991, Gudmundsson, 1997). Further, in Paper II we followed the position of Ingólfsson et al. (2010) where it is stated that a very quick deglaciation occurred as a response to a sea level rise which caused the largely marine-based ice-sheet to collapse. Since this relict population of rock glaciers is old and probably formed during the deglaciation of the late-Pleistocene glaciation, some considerations concerning the extent of the last glacial maximum (LGM) and deglaciation seem justified. First, rock glaciers are the cumulative expression of creep over millennia, and therefore a long and continuously ice-free time period is required for their formation. Second, a large-volume LGM ice-sheet is disputed by some scientists based on ages obtained from  $^{14}\text{C}$ -dates of marine sediments off the northern Icelandic coast (e.g. Andrews et al., 2000, Andrews and Helgadóttir, 2003, Van Vliet-Lanoë et al., 2007). Furthermore, according to Van Vliet-Lanoë et al. (2007) the LGM is northerly constrained mainly by assumptions of the presence of undated submarine terminal moraines from other parts of the island, i.e. far west and bordering the Reykjanes peninsula in southwest (Olafsdóttir, 1975, Ingólfsson et al., 1997). Third, the current understanding of the Younger Dryas advance in the Tröllaskagi is poorly spatially constrained in the most updated publications (e.g. Ingólfsson et al., 2010), and also disputed by tephra chronologies in north (Van Vliet-Lanoë et al., 2007) and by sedimentological analyses in the central south of Iceland (Geirsdóttir et al., 1997). Fourth, a climate in favour of rapid deglaciation is not consistent with a severe permafrost climate producing periglacial talus-derived rock glaciers at sea level. However, in the reasoning of Paper II it was argued that if the deglaciation was mainly a dynamic response to a rising sea level the climate could still be severe. Although the sea level rose within this chronozone, and perhaps triggered by



an Antarctic meltwater pulse (e.g. Weaver et al., 2003), the Bølling interstadial is associated with distinct a temperature rise as interpreted from ice-core records (Dansgaard et al., 1993, Stuiver et al., 1995), dendrochronology (Friedrich et al., 2001), and thermohaline circulation patterns (Ruhlemann et al., 1999, Schmidt et al., 2004). In this context, onset of rock glacier formation during the warm Bølling interstadial appears unlikely. Based on these considerations I suggest that rock glaciers in Iceland formed during a less voluminous LGM than commonly accepted, and/or since a pre-Bølling deglaciation. Here, a ‘minimum’ Younger Dryas ice-sheet model is assumed for Tröllaskagi, with the possibility of local glaciers forming in its surroundings.

### **5.3 Holocene permafrost distribution and permafrost age**

The simulation of permafrost occurrence over the Holocene both in time and space provides new insights to understanding the current pattern of permafrost in the Norwegian mountains. An important result in this respect is that permafrost survived the HTM at high altitudes in southern Norway, above c. 1800 m a.s.l. The temperature series used to drive the simulation, however, is somewhat conservative, and higher temperatures at the HTM might have occurred. If so, the zone of permafrost survival would be shifted towards higher altitudes, although lower than the highest summits. The CryoGRID1.0 model predicts permafrost occurring at c. 6.4 % of the Norwegian landmass at present (1981-2010). The last normal period (1961-1990) was colder, and permafrost is predicted for c. 10 % of the landmass (Gisnås, 2011, Gisnås et al., *submitted*), whereas the respective numbers for the HTM and LIA are 1.1 % and 14.5 %. Thus, the LIA permafrost extent was more than doubled compared to the last normal period, although the contemporary larger glacier area is not accounted for, and the permafrost distribution over the Holocene appears highly dynamic (Figure 17).

The relative age of mountain permafrost is a new aspect within permafrost classification. Relative ages previously exist for North American and Eurasian Arctic permafrost, and a similar approach is here introduced for mountain regions. The highly heterogeneous nature of mountain permafrost makes it hard to provide robust altitudinal zones for each mountain region, but if the zones are considered indicative, new insights can be achieved.

Due to the generally gentle relief characterizing the Norwegian permafrost regions these are expected to respond differently to climatic changes than for example permafrost in the European Alps. 3D-effects are an issue, even though vast areas of the landscape are characterized by plains. Repeated local and regional glaciations have incised these plains, and based on the properties of the glaciation both peaks and less affected high-altitude plains are common features. Large areas are underlain by permafrost, making the regions sensitive to temperature changes. If the lower limit of mountain permafrost in flat areas rises, large areas will be affected, analogous to a rise of the ELA in plateau-type glaciers. This is exemplified by for example the lower limit of HTM permafrost that rose above the southern Norwegian plateaus leading to permafrost prevalence only at the highest summits, or by the vast reduction of permafrost since the LIA in the Finnmarksvidda plateau in northern Norway.

Permafrost stabilizes its host material, both unconsolidated material and bedrock (Gruber and Haeberli, 2007, Etzelmüller and Frauenfelder, 2009). The highly dynamic nature of the permafrost presence over the Holocene must have inferred destabilization of slopes in periods of permafrost degradation, such as leading into the HTM and from the LIA until present. However, the slope stability over the Holocene is rather contrary documented to

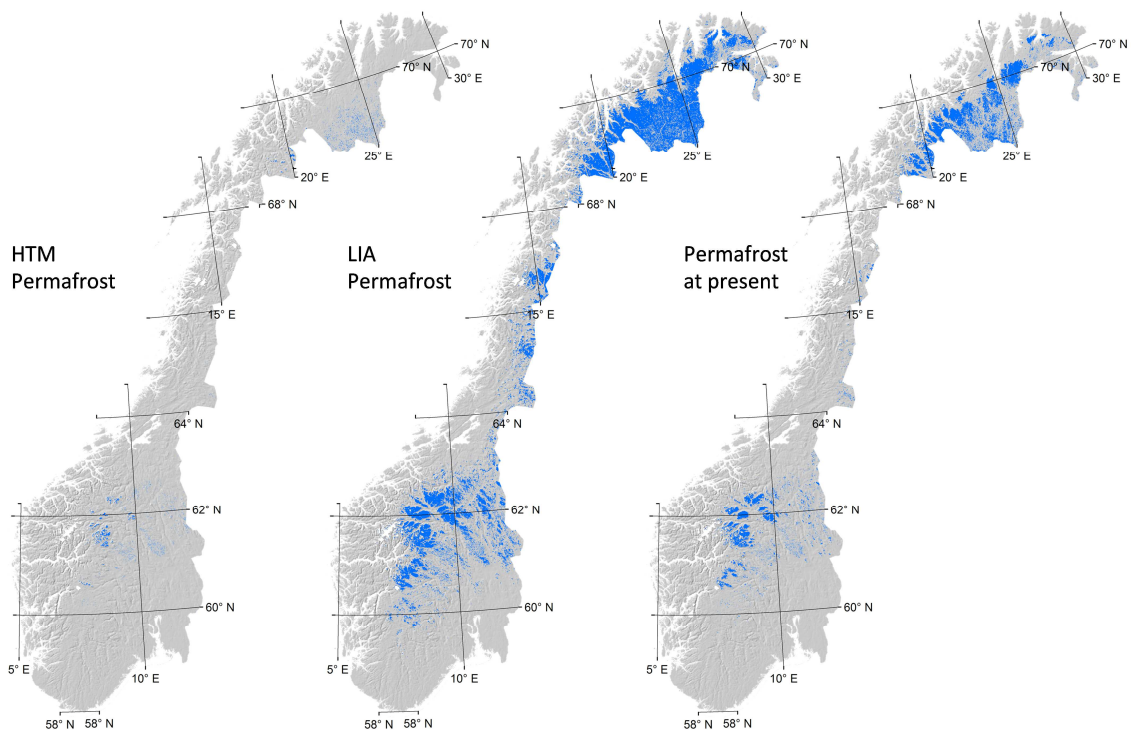


Figure 17 Permafrost distribution at (from left) the Holocene thermal maximum, the 'Little Ice Age' and the last normal period (1961-90).

decrease in periods of modelled permafrost aggradation, especially following the HTM at the onset of the Neoglaciation and onwards (Blikra and Nemec, 1998, Sletten and Blikra, 2007). It should be noted that these colluviums often, but not always, are situated lower in the terrain than the permafrost and might have been induced by pluvial processes rather than frost instability.

## **5.4 Cryoconditioned landscape evolution**

### *5.4.1 Holocene*

Geomorphologically, glacier-permafrost interactions are imprinted in the presence of landforms like open-system pingos, moraine-derived rock glaciers, and ice-cored and push moraines (Boulton, 1972, Liestøl, 1977, Benn and Evans, 1998, Lyså and Lønne, 2001, Etzelmüller and Hagen, 2005). In mainland Norway and Iceland, moraine-derived rock glaciers and ice-cored moraines occur frequently in certain regions, a sign of glacier-permafrost interactions. Further, substantial parts of what constitutes the paleic surface of Norway are covered in blockfields. Since the blockfields represent a ground thermal anomaly, these regions will first develop or maintain permafrost if the region are undergoing a climate deterioration or is at marginal permafrost conditions. This ground thermal anomaly caused permafrost to grow and persist during the build-up phases of the Pleistocene glaciations, and certainly in the surroundings of the varying ice-cover within one glacial cycle.

If considering contemporary glacier-permafrost interplay in a west-east transect of southern Norway, the different pro-glacial landforms produced carry ground thermal significance with them. Whereas the MPA decrease towards east, the ELA on glaciers increase (Etzelmüller et al., 2003). In the western areas, the ELA is situated below the MPA, temperate glaciers terminate in non-permafrost environments and ‘bulldozing’ moraines are formed. More eastwards, for example in the Finse region (Paper IV), the ELA and MPA are situated at approximately the same altitudes, and push-moraines are produced by cold glacier margins in polythermal glaciers existing in a sporadic permafrost environment. In areas of continuous permafrost, for example in Jotunheimen, the MPA is situated below the ELA, and glaciers are generally cold, either completely if small or more often partly, which is

geomorphologically manifested as well-developed ice-cored moraines and moraine-derived rock glaciers.

Rock glacier research has been a major focus within the discipline of periglacial geomorphology, but has also led to confusion originating from the discussion of glacial or periglacial origin. As most scientists now agree that rock glaciers belong to permafrost environments, it appears to be harder to agree on periglacial or glacial origin. Several processes may lead to the typical appearance of a rock glacier; however, the common denominator is the primary or secondary gravitational creep of unconsolidated debris caused by permafrost presence. Whether the creeping debris is of glacial origin (e.g. moraine, till or other proglacial deposits) or of periglacial origin (e.g. talus slopes, colluvium or other weathering products), is of secondary importance in this context. Still, the moraine-derived rock glaciers and ice-cored moraines can serve as links between periglacial and glacial geomorphology.

#### 5.4.2 *Pleistocene*

The topography of Norway as a whole varies considerably, and partly depending on relief and surface conditions, the ground response to climate perturbations differ substantially (Etzelmüller et al., 2007b). For example, the low-elevated Finnmarksvidda plateau in northern Norway is covered in till which delay climate perturbations response in the ground, whereas high-altitude regions to a larger degree show sparse surface covers or are subjected to the negative ground temperature anomaly caused by blockfield covers (Balch, 1900, Harris and Pedersen, 1998, Juliussen and Humlum, 2008). Different concerns are associated with the permafrost landscape types.

First, paleic surfaces are found in northern Norway at relatively low elevations and are permafrost sensitive mainly because relatively small changes of the present MPA affect large areas (Paper III). However, the till-cover delays the temperature signal into the ground, causing prolonged ground response times to air temperature variations. Second, high-altitude paleic surfaces such as those widely occurring in southern Norway between Quaternary valleys are often covered by openwork blockfields (Figure 18). The blockfields themselves represent a negative ground thermal anomaly as cold air that sinks in winter are trapped below lighter warm summer air closer to the surface. Thus, blockfields are also highly sensitive to permafrost, and, with the exception of palsa bogs, commonly represent the lowermost areas of permafrost. Third, in high-relief high-altitude mountain permafrost regions, the permafrost distribution is influenced by 3D-effects, and past cold periods such as

the last ice age still affect mountain permafrost temperatures due to latent heat release and lateral heat fluxes (Noetzli and Gruber, 2009). Even if ground temperatures below the depth of zero annual amplitude (ZAA) are stable, the transient effects of past climate modify the temperature fields within the mountains (Lunardini, 1996, Kukkonen and Šafanda, 2001). Sparse surface covers are common in such regions, and near-surface response to air temperature changes occur rapidly.

In this respect, blockfields represent a very interesting element of the Norwegian landscape. The blockfields survived the Pleistocene glaciations at coastal palaeo-nunataks (Nesje, 1989, Nesje and Dahl, 1990, Brook et al., 1996, Rea et al., 1996), such as in northwestern parts of southern Norway and in the Lyngen-region of northern Norway (Sollid and Sørbel, 1979), and where it is evident by the occurrence of erratics, lateral melt-water channels and moraine-deposits below cold-based glaciers (Follestad, 1990, Sollid and Sørbel, 1994, Fjellanger et al., 2006).

During the Pleistocene glaciations large parts of the landscape were covered by cold-based ice which preserved rather than eroded its substrata (Kleman, 1994, Sollid and Sørbel, 1994, Kleman and Stroeven, 1997, Fjellanger and Sørbel, 2007, Kleman et al., 2008). In Tertiary the landscape was characterized by fluvial drainage and erosion, and when the Quaternary ice-sheets built up the natural drainage paths were through pre-existing fluvial valleys (Gjessing, 1967, Nesje and Whillans, 1994). Before substantial ice thicknesses were accumulated, the glacial erosion would work most efficiently where the ice velocity was highest, i.e. through the fluvial valleys, and this initial elevation difference and focused erosion would quickly develop into a positive feedback loop. As the glaciers grew and got more erosive, the ice drainage patterns were largely set through the pre-existing and increasingly deeper and wider valleys, which could drain increasingly more ice. Current ice flow observations in Greenland and Antarctica (Joughin et al., 2010, Rignot et al., 2011) show patterns of high ice velocities concentrated in ice-streams and separated by large areas of low-velocity ice, which probably also reflect the drainage patterns of the Pleistocene glaciations. Slow ice flow and relatively thin ice covers at high elevations in combination with low air temperatures caused cold ice to persist, and the paleic surfaces and corresponding blockfields was protected from erosion. From present conditions, it is known that blockfields represent areas where permafrost is likely to form rapidly during climatic deteriorations, as for example during the onset of the Quaternary ice ages, as well as during

interstadials within the glacials. As ice-sheets require millennia to grow, once they aggraded over blockfields at high altitudes the glaciers were affected by the underlying ground thermal anomaly. As a combined result of shallow ice and initially frozen ground, blockfields were protected through the repeated glaciations. The blockfield evolution model by Ballantyne (2010) suggests that the present-day blockfields are not completely preglacial phenomena, but rather the resulting effect of Tertiary chemical weathering forming saprolites and corestones, processes which during the Pleistocene were replaced by removal of fine-grained material, frost wedging and vertical frost sorting. Active rock jointing at the interphase between bedrock and blockfields were governed by the presence of permafrost and the blockfield thickness, although lowering of surfaces occurred, was equal to the active layer thickness (Ballantyne, 2010). Close and interacting relationships between blockfields, permafrost and cold-based glaciers preserved the preglacial paleic surfaces of Scandinavia and both protected and partly formed the blockfields (Figure 18).

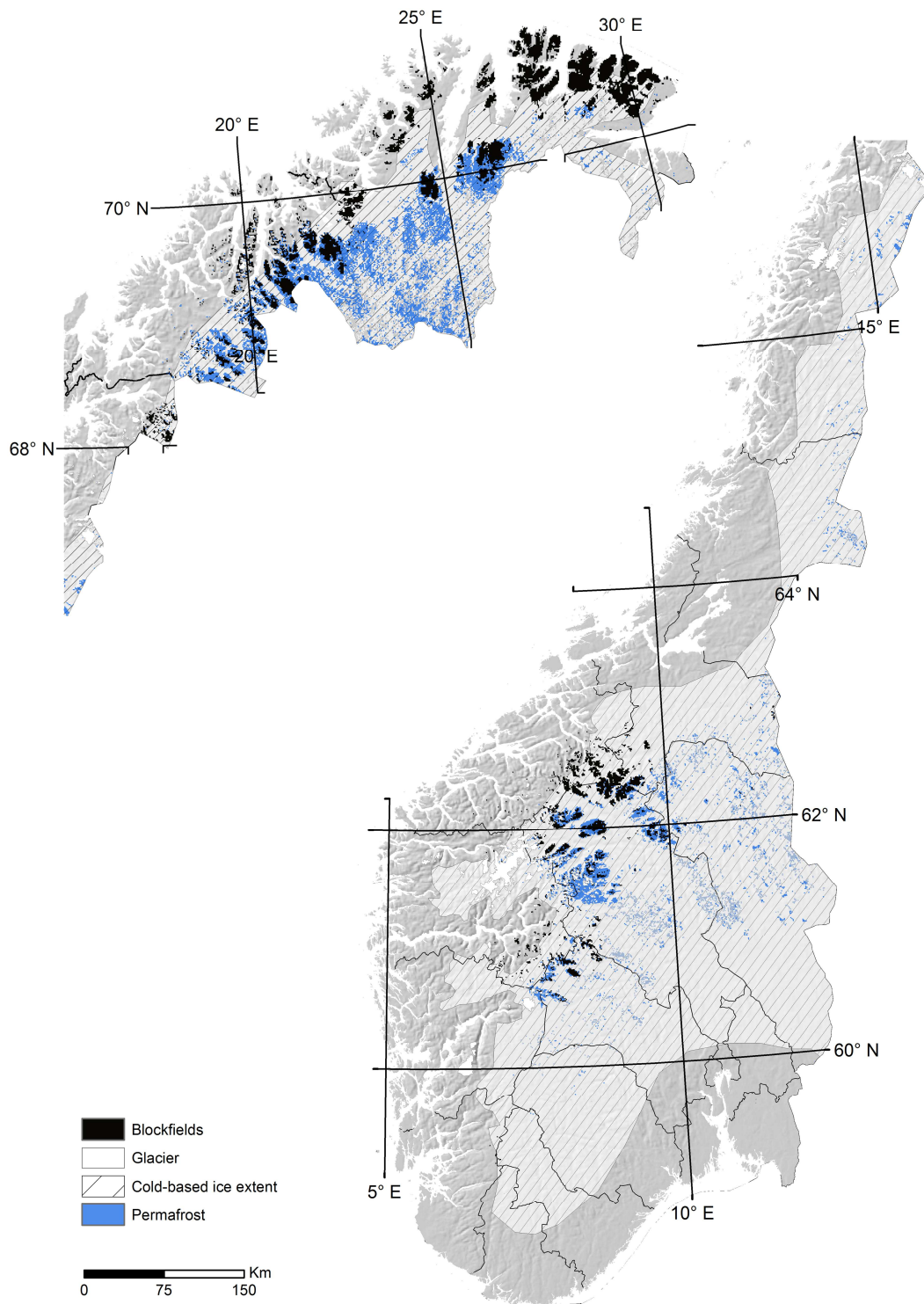


Figure 18 Illustration of areas covered by blockfields (classified by A. Ellingsgård, unpublished), present-day permafrost (Gisnås, 2011) and areas indicative of cold-based ice (Kleman and Hättestrand, 1999).

## Chapter 6

# Conclusions

In Norway, the general abundance of landforms is low compared to other mountain permafrost regions whereas the abundance is high in Iceland. The majority of the modern landforms in both regions are connected to glacial activity, expressed as landforms such as moraine-derived rock glaciers and ice-cored moraines. However, the relict landforms are mostly talus-derived rock glaciers. A shift in the dominating processes forming permafrost landforms occurred over the Holocene, from an early-Holocene dry, periglacial environment to a mid- to late-Holocene moist, glacial environment. Occurrence of relict rock glaciers at sea level in Iceland indicates a less extensive last glacial maximum or an earlier and faster deglaciation than the commonly accepted model predicts. The current landform mode in both regions is interpreted to represent maritime permafrost.

Modelling of 1D and 2D permafrost extent over the Holocene indicate that high-altitude permafrost in southern Norway is of pre-Holocene age and survived the Holocene thermal maximum (HTM), whereas the 'Little Ice Age' (LIA) permafrost area was more than doubled compared to present. Glacier and permafrost aggradation after the HTM increased the potential for glacier-permafrost interactions. Two periods of permafrost degradation were identified after the deglaciation, first at the onset of the HTM and second from the LIA maximum until present.

A small-scale case study of glacier-permafrost interactions from the currently very small glacier Omnsbreen in Southern Norway suggests that the lack of glacier-marginal landforms from the maximum extent may indicate a formerly cold glacier margin. This



glacier formed at the onset of the LIA at c. 1425 AD, largely disappeared during the 19<sup>th</sup> century, and the spatial permafrost model (CryoGRID1.0) suggests continuous permafrost in the area during its formation. At present the glacier depends on wind-redistributed snow to exist, whereas sporadic permafrost in the area exists at wind-blown sites.

Also during the build-up phases of the last glaciations, permafrost was likely to be present and to affect the subglacial thermal regime. Especially in areas of blockfields a close interaction between subglacial thermal regime and ground thermal properties caused preservation of the landscape.

Together, the current geomorphic expression of the landscapes in Norway and northern Iceland are affected by a long-term cryoconditioned evolution.

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## **Part II Papers**





# Paper I



# Paper II



# Paper III



# **Paper IV**

