Cenozoic landscape development on the passive margin of northern Scandinavia

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The geometry of major bedrock landforms in a central part of the Northern Scandes has been examined for interpretation of landforming events and their chronology. Analysis of palaeosurfaces and palaeovalleys was undertaken based on digital elevation data. Four major landscape generations, governed by different base levels, were reconstructed within the mountains. In the east the generations more or less merge and form plains with residual hills. The configuration of the landscape generations suggests an asymmetrical uplift with maximum uplift in the west caused by discrete events during the Cenozoic and guided by reactivated Mesozoic faults west of the study area. The style and amount of uplift differs from the Southern Scandes and a hinge line is defined in between. The landscape generations were also used for estimations of the amount of glacial/fluvial erosion since the beginning of the Late Cenozoic glaciations. Erosion amounted up to 200 - 400 m in the valleys that hosted the major outlet glaciers, while the eastern plateaux and the plains with residual hills have experienced no or limited glacial erosion.

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Introduction

Preserved base-level governed palaeosurfaces occur in formerly glaciated areas and have been shown to be useful for estimation of both the magnitude and style of uplift events (Lidmar-Bergström et al. 2000, Bonow et al. 2006a, 2006b). Glacial erosion was long supposed to have been a major agent of erosion in formerly glaciated mountain areas, but it is now well established that the late Cenozoic subglacial erosion has been highly variable (Sugden 1978; Kleman & Stroeven 1997; Staiger et al. 2005; Li et al. 2005; Harbor et al. 2006). Landscape generations in stepped sequences were identified at an early stage in northern Scandinavia (e.g. Wråk 1908; Rudberg 1954). The identification was based on maps with poor accuracy of height data and registered in combination with field work. We have tested these observations and now consider their importance in relation to the timing and amount of Cenozoic uplift and passive margin development of northern Scandinavia (Jensen & Schmidt 1992; Riis 1996; Lidmar-Bergström 1999; Lidmar-Bergström & Näslund 2002; Faleide et al. 2002; Hendriks & Andriessen 2002; Mosar 2003; Redfield et al. 2005). In this study we use digital elevation data to make a reconstruction of palaeosurfaces and palaeovalleys along closely spaced profiles in an area of the Northern Scandes and their eastern flank. The reconstruction of fluvial landscape generations also allows estimation of erosion, particularly glacial erosion, after formation of the pre-glacial landscape. Finally, we use the results to compare the style and timing of uplift between the Northern and Southern Scandes.

The study area

The study area is located in northern Scandinavia and extends from the Norwegian coast in the west across the NNE-SSW striking elevation axis of the Northern Scandes to the plains in the east (Fig. 1). The broad valleys with the Torneträsk and Akkajaure lakes define the northern and southern boundaries. The water divide is northwest of the elevation axis and also northwest of the study area in the two main valleys. The area can be divided into four morphological regions: 1) The western highlands (west of the main elevation axis) have some high mountains separated by low areas and wide valleys. 2) A high mountain zone consists of a group of high mountains, with Kebeke lakes as the highest summit, 2097 m a.s.l. (summit glacier excluded). The highest areas of regions 1 and 2 are often characterized by alpine relief (a fretted upland with peaks, horns and arêtes) and numerous small valley glaciers. 3) Low mountains to the east with flat surfaces at 800 and 1000 m a.s.l., between the incised valleys draining eastwards. 4) The eastern plains with inselbergs rising at the most 400 m above the surrounding landscape (Dundret, summit 821 m a.s.l.).

Bedrock, structure, and topography

The bedrock consists of Caledonian nappes in the west and Precambrian basement rocks in the east (Fig. 2). The mountains, here defined as areas rising above 800 m a.s.l., extend from the coast in the west to 10 - 40 km
east of the Caledonian front (Fig. 1). Four major bedrock units make up the mountains within the Caledonian rocks (Gee et al. 1985). The Seve nappe (part of the upper nappes) forms the highest mountains. The overlying Koli nappe follows westwards and also a large area with Precambrian rocks, as a part of the Caledonian nappes. The middle nappes are exposed in the south and north and also along the Caledonian front. Autochthonous or parautochthonous Vendian and Lower Palaeozoic rocks follow eastwards. They rest on the Precambrian basement, which underlies the eastern part of the study area. The direction of the main valleys is to a large extent structurally controlled. The NE-SW trending major tributary valleys are often located at the contact between the different nappes.

**Denudation history**

Apatite Fission Track (AFT) analyses of samples from the Northern Scandes have revealed a period of continuous erosion since the Carboniferous, and onset of a final erosional period in the Late Cretaceous – Paleogene (Hendriks & Andriessen 2002, Hendriks 2003). These data further show that 1 – 2 km of rock was eroded from the region during the Triassic and Jurassic and 1.2 – 2.2 km of rock in the Cretaceous. Cenozoic erosion was estimated to between 0 and 3 km. Thus nothing in the present relief in the Northern Scandes ought to be older than the Late Cretaceous unless exhumed from cover rocks.

Analysis of landforms just east of the study area suggests the existence of preserved sub-Cambrian facets (Rudberg...
1988, Lidmar-Bergström 1996) and thus the former existence of a Palaeozoic cover. A sedimentary cover is modelled in several studies and it is suggested this is caused by erosion of the rising Caledonides (Tullborg et al. 1995; Larsson et al. 1999; Cederbom et al. 2000; Huigen & Andriessen 2004), but the thickness and extent is controversial (Hendriks & Redfield 2005; Green et al. 2006). Erosion of this cover in the study area must have started in the Carboniferous/Permian (Hendriks 2003). After this erosional event the basement of the study area was probably re-exposed and subjected to deep weathering, as is shown by preserved kaolinisation (Frietsch 1960, refs. in Elvhage & Lidmar-Bergström 1987; Lidmar-Bergström 1995).

Wråk (1908) noted that the water divide was located northwest of the whole study area when the valley systems developed. He identified remnants of two generations of base-level-governed erosion surfaces (Tuipal and Borsu) dipping ESE all the way from the Norwegian coast in the west towards the mountain front in the east (Fig. 3). Minor remnants of one or two older surfaces, Likka, were identified above and a valley generation, incised in the Borsu surface that continued into the eastern plains (the Mud dus plains). On the western side of the mountains a correlative generation dipping towards the NW was identified (the Guris generation). Two lower valley generations were identified in the peripheries but are not dealt with here.

Glacial erosion of the area has occurred during three different glaciological regimes (Kleman & Stroeven 1997):

Fig. 2: Bedrock map, compiled from Metamorphic, structural and isotopic age map, Northern Fennoscandia, 1:1 mill. (Geological Surveys of Finland, Norway and Sweden 1988) and Gee et al. (1985). Bands a – e show location of topographical profiles in Fig. 6.
by cirque glaciations during interglacials, by mountain-centred ice sheets (Jansson & Fredin 2002), and by east-centred Fennoscandian ice sheets. Linear glacial erosion changes old fluvial valleys by deepening and widening them with the formation of U-shaped valleys (Montgomery 2002; Bonow et al. 2003). In this way local base levels can be lowered and therefore fluvial incision may be enhanced during interglacial periods (Wråk 1908, Rudberg 1992, 1993; Kleman & Stroeven 1997). Unlike fluvial erosion, glacial erosion is not controlled by any base level, and can for example erode well below sea level along valleys fed by several merging valley glaciers (e.g. over 1000 m in the Sognefjord) (Nesje & Whillans 1994). On the eastern flank of the Northern Scandes, erosion in marginal lake basins has been less than in the west, e.g. in Hornavan (220 m deep) (Lindkvist & Danielsson 1987). Areas situated at high elevations, such as inter-valley flats, often experienced cold-based conditions during glaciations and are thus only marginally altered by glacial erosion (Kleman & Stroeven 1997; Stroeven et al. 2006).

**Methods**

**Landform analysis**

In this paper we deal with erosion surfaces in stepped sequences and incised valleys, formed in a valley-in-valley
system (Rudberg 1954, Ahnert 1998 p. 225-227, Bonow et al. 2003). An erosion surface in this paper is defined as a surface of low relief constrained by a common base level for the fluvial system (Bonow et al. 2006a). Such surfaces can form through valley widening at the same time as the valley heads retreat by backward fluvial erosion (Ahnert 1998 p. 218-219). The widening of valleys is caused by slope retreat from the centre of the valley forming rock terraces (pediments). Uplift causes incision and initiates the formation of a new valley within the old one leaving surfaces (valley benches) along the sides (Ahnert 1998 p. 183-184). The sequence of surfaces and valleys formed makes up stepped surface landscapes or landscape generations.

Pediments mainly have slopes between 4º and 6º with a maximum of 11º (Tabor 1952; Dohrenwend 1994). Bonow et al. (2003) made a best fit analysis of slope angles for surface mapping in an area in south central Norway with both Precambrian and Caledonian rocks and found that the slope angle 6.5º gave the best result for separating surfaces from slopes. In this paper we have used this value of angle for tracing palaeosurface remnants.

A fluvial valley is broad at its mouth and narrow in the inner part. In dendritic valley patterns the angle with which the tributaries join the main valley also shows its direction of formation if not influenced by structure (Ollier 1981; Summerfield 1991). We used these conditions for construction of a palaeodrainage map to aid interpretation of the fluvial landscape development.

Analysis of elevation data

A Digital Elevation Model (DEM) with a 50 m resolution (the National Swedish elevation database) was used to analyse the topography of the study area in a Geographical Information System. A height interval map was produced to get an overview of valley patterns and general topography (Fig. 1). On the basis of a simplified height layer map
Fig. 4a), the palaeodrainage system was reconstructed (Fig. 4b). Detailed topographic maps were consulted for evaluation of glacial rearrangements of the preglacial fluvial pattern. Identification of preserved landscape generations was performed by analysis of a surface/slope map (Fig. 5) in combination with topographical profiles (Fig. 6). The surface/slope map was constructed as follows. Slope angles were calculated for each grid cell. Areas were then classified as surfaces if the slope gradient for each cell was < 6.5° and as slopes if the gradient was higher. The surfaces were coloured according to height above sea level. Height classes were chosen to fit in between slopes as well as possible along the mountain front in the east, where a stepped landscape is best preserved. In the west the landscape is more dissected and it was difficult to correlate remnants of a stepped surface landscape without detailed analysis. Therefore height profiles at the same scale as the map were used for the combined analysis (Fig. 6). Nine profiles, N60ºW-S60ºE, perpendicular to the main elevation axis, were constructed across the study area. Profiles for maximum and minimum heights in a 5 km wide zone along each profile were added. Summit surfaces, plateaux below summits, and valley benches from these zones on the surface/slope map were marked manually on the profiles. Correlation of surface remnants was made within each profile and also between nearby profiles by following the palaeodrainage.

A regional profile was constructed covering a 130 km wide corridor, including the study area, from Lofoten in the northwest to Finland, north of the Gulf of Bothnia, in the southeast. Maximum and minimum values for each cross section along the corridor were plotted together with a central profile. The diagram was constructed from a DEM of Scandinavia with a 500 m resolution (according to Lidmar-Bergström & Näslund 2002 with data from Swedish National Land Survey and Statens Kartverk, Norway). A summary of the landscape generations was marked in the profile.

Results

Fluvial dissection

The trunk valleys out from the mountains are the Torneträsk/Torne älv valley and the Akkajaure/Lule älv valley. Between these valleys four palaeodrainage areas were identified (Fig. 4a, b) as follows:

1) Valleys with a southeasterly direction developed along the eastern flank of the mountains.

2) Northwest of the high mountain belt, a major tributary to the Torneträsk valley, the Abisko valley with its tributary valley Alip, has penetrated far to the southwest. A smaller tributary valley to the Torneträsk valley further east only penetrated to the southwest for a short distance. These two valleys have broad open mouths to the Torneträsk valley.

3) Two major tributary valleys to the Akkajaure valley penetrate far to the northeast and north forming the Sitasjaure/Ritsem valley and the Tjäktjavagge/Vakkota-
vare valley. Their broad open mouths, inward narrowing and the direction of their minor tributaries demonstrate their fluvial origin and the original direction of water flow. The river in the Tjäktjavagge/Vakkotavare valley must have abandoned its distal portion (the present position is too high) before the Late Cenozoic glaciations and has probably been captured by northwestward drainage. Here is a broad and open valley, while the valley heading southeastwards with the present drainage is a narrow glacial gap (Kleman & Stroeven 1997).

4) A palaeodrainage area in the west has two deeply incised valley systems directed towards Rombaksfjorden and the Skjomen fjord. A high level valley with a westerly direction is identified. It is related to the southern valley system. Denudation within palaeodrainage areas 2 and 3 has formed the main part of the preglacial landscape within the western highlands.

Surfaces and slopes
Analysis of the surface/slope map (Fig. 5) gave the following result. The mountains rise between the tributary valleys with typical plateaux sloping towards the valleys. Generally the highest surfaces form isolated remnants on or around mountain summits. Large areas with low relief can be followed all around the main elevation axis between 1000 m and 850 m a.s.l. Wide and shallow valleys occur at 850 – 650 m a.s.l at the edge of the mountains in the east. Below this level the Muddus plains expand east of the mountains. Flat areas in low positions

Fig. 7: A regional profile across the Northern Scandes (NS) and its eastern flank with maximum and minimum values extracted from a 130 km wide zone along the profile. Four fluvial surface/valley generations dip towards ESE all the way from the Norwegian coast. They join more or less in the east with generations 3 and 4 making up the floor of the Muddus plains. Faults in the eastern Vestfjorden, and east of Vågsfjorden and Ofotsfjorden are interpreted to have guided an uplift centre close to the coast. The escarpment at the coast is 500 m high from generation 4b. The bottom profile shows the lake levels of the glacially eroded main valleys. The step in the bottom profile in the west shows the backward incision of the fjord valleys. F = fault. Ve = Vestfjorden. O = Ofotsfjorden. Vå = Vågsfjorden. Ke = Kebnekaise. D = Dundret.
within the mountains (e.g. valley flats) mainly reflect surfaces of glacial erosion. This is evident from the reconstruction of the fluvial generations (see below).

**Major fluvial generations**

The fluvial landscape development was analyzed by the aid of the profiles. Five of the nine profiles were chosen for presentation (Fig. 6). Four major generations with an overall inclination towards ESE were identified in all profiles and in addition a summit generation (0) of limited extent. They correspond, within the mountains, to those of Wråk (1908), viz. 0 = Likka, 2 = Tuipal, 3 = Borsu, and 4 = Muddus. The inclination of the landscape generations towards the east is interrupted by the major tributary valleys, which have been decisive for the successive lowering of their surroundings (cf. Fig. 5). Remnants of generation 0 occur mainly within the Seve nappe (profiles c – e). Generations 1 and 2 make up the summit surfaces with major surfaces preserved at the mountain front in the east. The surfaces of generations 2 and 3 cut across the Caledonian front. Major valleys, representing generation 4, are incised in surfaces belonging to generation 3 within the mountains. East of the mountains generations 3 and 4 merge to form the Muddus plains. These four landscape generations have been eroded by eastward directed drainage.

A generation 4a is eroded by rivers with a westward drainage across the area covered by the western part of profile d. A lower generation (4b) is also eroded by westward drainage across a larger area (profiles b, c and d). Valleys are deeply incised in this generation and make up an escarpment in the west about 500 m high.

**Exhumed sub-Cambrian facets?**

Profiles a and c (Fig. 6) indicate that sub-Cambrian facets might occur in the eastern parts of the profiles as the summits occur along a straight line and the maximum, minimum and middle profiles almost coincide.

**The regional profile with a summary of landscape generations**

The five generations are summarized in the regional profile (Fig. 7). The surfaces are about 300 and 400 m apart in the west but in the east they more or less merge with generations 3 and 4 making up the bottom level of the Muddus plains. The lowest generation approaches horizontality while the top surfaces dip most. The summit of Dundret might be correlated with generation 0 within the Seve nappes. The eastern part of the profile continues outside the study area. Here summits within individual tilted bedrock blocks can belong to remnants of the sub-Cambrian surface (Lidmar-Bergström 1996). The bottom profile shows the lake levels of the glacially eroded main valleys. The step in the bottom profile in the west shows the backward incision of the fjord valleys.

Landscape generations 1 – 4 are developed by eastward drainage and with a water divide located west of the study area. The drop in height in the maximum profile on the western side of generation 0 is caused by denudation connected to eastward drainage of generation 2. Generations 4a and 4b are developed by westward drainage. The present escarpment (500 m high) is developed by incision in level 4b of westward drainage.

**Estimation of glacial erosion.**

The plateaux of the low mountains east of the elevation axis are examples of well preserved palaeosurfaces. The extent and form of these plateaux give constraints on the aerial extent of deep glacial erosion. The surface/slope map (Fig. 5) shows that the plateaux at 850 – 650 m a.s.l. south of eastern Torneträsk and east of Akkajaure have been partially destroyed by substantial glacial erosion, while they are better preserved between these areas.

The valleys penetrating westwards into the mountain front are glacially reshaped to different degrees. The identification of preglacial landscape generations (Fig. 6) gives constraining levels from which later erosion was estimated along the profiles. It is difficult to judge how far the different generations, particularly generation 4, had penetrated into the mountains before the glaciations started. However, the minimum profiles clearly show where there has been substantial glacial erosion even below this generation. The glacial erosion at the mountain front (profile a) has cut into generation 3 and removed up to 250 m of rock, but 20 km from the mountain front the present relief coincides with generation 4 and the glacial erosion is estimated to be minimal. Rautasjärvi valley penetrates at a low level far into the mountains. Erosion below the preglacial level calculated from generation 3 amounts to 450 m in the inner parts and to about 250 m at the mountain front. A similar value (250 m) is estimated for erosion below generation 4 in the Abisko valley (profiles a and b). Paitasjärvi valley (profile c) has been lowered about 250 m and Teusajaure (profile e) more than 300 m below generation 4 by glacial erosion. At the mountain front glacial erosion along profile e amounts up to 320 or 160 m depending on whether generation 2 or 3 is the correct reference surface. Glacial erosion was estimated to have been up to 250 m below generation 4 in the western part of profile e. The depth of the lakes has to be added to get the maximum depth of glacial erosion. This gives a total depth of 390 m for the glacial erosion of Teusajaure valley (max depth of Teusajaure lake is 71 m; Lindqvist & Danielsson, 1987). Thus pronounced glacial erosion has only occurred in restricted areas.

**Discussion and interpretation**

**Reflections on constructions of landscape generations**

In small and less dissected areas with no or only minor glacial erosion, as is the case along the upper parts of
Gudbrandsdalen in south Norway, mapping of preglacial erosion surfaces is easier than in the present study area (Bonow et al. 2003). Profiles were needed for correlation as much rock has been removed by glacial erosion in our area. The palaeosurface/palaeovalley remnants marked on the profiles gave a regular pattern with inclined palaeosurfaces that appeared in all the nine profiles. These surface generations are only presented in the profiles. The generations form a sequence of inclined surfaces within the mountains. We were able to ascertain the larger difference in height between the surfaces in the west and the merging of generations 3 and 4 outside the mountains in the Muddus plains in the east. Wråk (1908) did not note the mergence of the surfaces and in other reconstructions the surfaces are often described as more or less horizontal (e.g. Rudberg 1954 p.406).

**Relative age of landscape generations and the uplift of the Northern Scandes**

The described sequence of inclined surface and valley generations in the study area has most likely been formed by fluvial incision and valley widening as a result of uplift events. The landscape generations have developed along the valleys at the expense of the generation above and thus the highest generation is the oldest. The relative vastness of each surface indicates the relative time available for its formation before further development was interrupted by an uplift event (Wråk 1908, Bonow et al. 2006a). The landscape generations can be used as event markers (Schoenbohm et al. 2004, Bonow et al. 2006b).

The time available for the final shaping of the Muddus plains involves time for the formation of both generations.
3 and 4. The residual hills of the plains might have a partial origin in an irregular Mesozoic etch surface (kaolinisation along fracture zones and soft ores; see Denudation history) preserved for a long time below a cover of Cretaceous sedimentary strata (Lidmar-Bergström 1995; Hendriks 2003). Uplift leading to erosion of the Mesozoic cover started in the Late Cretaceous (Hendriks 2003) and the cover must have disappeared already in the Eocene. Re-deposited marine Eocene diatomaceans are encountered in lake sediments south of the study area within the Muddus plains (Cleve-Euler 1941). This suggests a Palaeogene component in the present landscape, but the age for the final shaping of the Muddus plains could be much younger. The Late Miocene is a possible time for this with semiarid climates promoting pedimentation and erosion of old kaolinitic saprolites (Lidmar-Bergström 1982, p. 164, 172). The summit of the residual hill Dundret (profile e) belongs to generation 0. This high hill might have an origin in a Mesozoic etch surface and the cut off summit thus hints at a younger planation event in the Late Cretaceous/Early Palaeogene, constraining the age for generation 0. Based on our present knowledge, therefore, the possible time span for the formation of the stepped relief down to generation 4 seems to be from Late Cretaceous to Late Miocene. Thus only Cenozoic uplift events are registered in the present relief.

The landscape generations dip to the ESE, the upper ones more than the lower ones. The configuration of the generations (Fig. 7) suggests an asymmetrical uplift of northern Scandinavia with maximum uplift in the west. This conclusion is different from that of Rudberg (1954).
He argued for an ‘en bloc’ rise of northern Scandinavia along peripherally placed faults and flexures, a conclusion based on his mapping of more horizontal surfaces.

The step from generation 4b marks a latest uplift of about 500 m relative to sea level in the west. We interpret the abrupt knick in the bottom profiles (Figs. 6 and 7) to be part of the fluvial escarpment in the glacially eroded major valley floors caused by backward valley incision. Although broad and deep in the outer parts due to glacial erosion, these valleys are narrow and winding in their inner parts indicating continued fluvial action. This backward incision is thought to have occurred during interglacial stages and during melting of the ice sheets.

As most of the preglacial drainage from the study area (Fig. 4 b) transported debris eastward, where it is no longer preserved, the sedimentary record offshore in the west is of limited value for constraining landscape development before the glaciations. It is only material from the formation of generations 4 a and 4 b, lower fluvial incision, and glacial erosion that could be found offshore the coast of Norway, while material from the formation of generations 1 - 4 could have been stored in temporary covers in the east.

**Preservation and destruction of preglacial fluvial generations**

The uplifted surfaces are successively dissected by fluvial erosion and the highly elevated surfaces have been further eroded by cirque glaciers, particularly in the western areas (Fig. 8). Valley glaciers have eroded both valley sides and valley floors. Differences in the resistance of the bedrock have played an important role for how long a surface generation is preserved, but not for its formation (Bonow et al. 2003; Bonow et al. 2006a). For example, generation 0 is only preserved in the Seve nappes and in amphibolite (Dundrept). On the contrary, resistant rocks have not been decisive for the level of surface formation as sometimes is claimed (Summerfield 1991; Brown et al. 2000; Fjellanger & Etzelmüller 2003). The surfaces are regularly spaced regardless of bedrock, which supports this view. There is a continuous downwearing by weathering and surface wash of the plateaux. Probably there has also been a small amount of glacial erosion (cf. Stager et al. 2005; Phillips et al. 2006), maybe more in the west (cf. Kleman & Stroeven 1997). These conditions have not obliterated the sequence of steps, although the height of the steps relative to each other might locally have changed a little. Surfaces essentially untouched by glacial erosion are mainly present on the eastern low plateaux and on the highest plateaux (Kleman & Stroeven 1997). Low areas in the west are lowered by glacial erosion up to 250 m down to the surface of the major lakes. Specifically, some of the tributary valleys (e.g. Sitkasjaure), which broaden markedly westwards, were affected by ice moving westwards. We suggest that glacial scouring here was caused by bottom melting eroding ice due to the preglacially low position of the terrain. This in turn we think had been lowered by deep weathering and fluvial erosion in the easily denuded rocks of the Köli nappe. Despite massive glacial erosion westwards along the major valleys the present water divide here is still far to the west of the highest mountains.

The Muddus plains have experienced limited glacial erosion except where they are close to the former outlet glaciers of mountain-centred ice caps along the mountain front. Abundant remnants of gravelly saprolites below the till cover (e.g. Hirvas et al. 1988) indicate moderately deep weathering during ice-free periods of the Plio-Pleistocene, but this has probably only contributed marginally to the reshaping of the plains with residual hills.

**Landforms and passive margin development**

The study area is in a culminating part of a passive margin, the nature of which is under debate (Mosar 2003). Young margins of this type are often characterized by fault-related rift shoulders with steep topography and gentle back slopes (Weissel & Karner 1989). Mature margins can also show a high topography even if rifting started more than 250 million years ago as it did along the Norwegian margin. The origin of such topography has long been discussed (e.g. Peulvast 1985; Gilchrist & Summerfield 1994). There are two major models. One favours a marginal warp and fluvial incision resulting in a Great Escarpment (Ollier 1982) and the other a fault-related uplift followed by fluvial erosion and also the formation of a Great Escarpment (Persano et al. 2002). Both models assume a continuous high level margin since break up, though this is not the general view among Scandinavian geologists and geomorphologists concerning the Scandes. The Norwegian margin is well known (summary in Mosar 2003). The major rift events occurred in the Permian/Triassic, Late Jurassic/Early Cretaceous and Late Cretaceous/early Tertiary and the area of rifting migrated westward with time. Basins formed and were filled up with several km thick sedimentary successions. The rifting events were coeval with Mesozoic uplift periods resulting in deep erosion and the successive formation of major, low level erosion surfaces in southern Norway probably from the Late Cretaceous/Palaeogene into the Neogene, and now at high elevations (Peulvast 1985; Lidmar-Bergström et al. 2000). The exact correlation with erosional periods and sedimentation offshore is uncertain.

Mosar (2003) looks upon the Scandes as the geomorphic expression of uplift along an important extensional fault, the innermost boundary fault system (IBF). It runs from the Hardangerfjord in the south, through the southern Scandes and further northwards west of the basement windows within the Caledonian nappes and is mainly west of the topographical crest. Movements along this fault system in southern Norway are only ascertained until the Early Cretaceous. As no fault scarps are seen in the local topography in southern Norway, we conclude that the surface of the present high level plateaux must
be younger. Although these faults have been active during Late Palaeozoic/Mesozoic phases of rifting they do not therefore seem to be related to the present day configuration of the Southern Scandes. Instead a slight doming in combination with reactivation of faults, particularly along the Møre-Trøndelag fault zone (Redfield et al. 2005), seems to have guided the formation of surfaces since maybe the Late Cretaceous within the Southern Scandes (cf. Lidmar-Bergström et al. 2000). Similarly, no signs of fault scarps are visible in the landforms of the study area and it is suggested that the uplift here was controlled by fault systems along and east of the offshore basins (Fig. 7), although the uplift is dome-like along the extension of the mountain chain.

In the study area, the elevation axis of the Northern Scandes is due to resistant rocks and lies east of the water divide. The position of the present water divide is due first to the establishment of the westward drainage, and second to westward reversal and capturing of some rivers, as for example the upper part of the Abisko and Alip rivers (Fig. 4a, b), maybe during the glaciations. During formation of generations 1 – 3 the water divide and uplift centre have been west of the study area, and thus still further west than today. The onset of eastward drainage was probably a consequence of uplift in connection with the formation of a passive margin in the Late Cretaceous – Palaeogene, which culminated with sea-floor spreading in the Eocene (Skogseid et al. 2000). Wråk (1908) argued that there must have been land in the west when the highest surfaces in the study area developed. Lofoten was not above sea level in the Late Cretaceous/Palaeogene (Leseth & Tveten 1996; Brekke 2000; Tsikalas 2001). It can be imagined that the highest surface, generation 0, developed across an area with sedimentary strata in the west, Caledonian basement rocks, and over a cover of Cretaceous strata east of the present mountains. Later uplift events caused erosion of the covers and formation of the lower generations. Westward drainage developed successively in the easily eroded sedimentary sequence. First with the formation of generations 4a and 4b, new erosion surfaces were developed in Caledonian basement rocks along the western part of the study area. The Permo/Triassic and Permo/Jurassic basins offshore Mid-Norway are separated from the onshore area by the Border Fault Complex (Mosar 2003). This complex continues along the eastern side of the Vestfjorden Basin as the Hamarøya fault (Olesen et al. 2002). We suggest that some of these faults, partly offshore in Vestfjorden and Vågsfjorden and partly onshore, have guided a Cenozoic uplift centre close to the coast (Fig. 7). This does not mean that the uplift along faults has been of the same amount everywhere along the Scandes, but rather that there is greater asymmetry in certain settings both along the Northern and the Southern Scandes due to reactivation of old faults.

### Correlation of landscape generations between Northern and Southern Scandes and comparison of timing and style of uplift

The geometry of the surface generations in the study area differs from those of the Southern Scandes (Lidmar-Bergström et al. 2000b; Bonow et al. 2003; Fjellanger & Etzelmüller 2003) (Fig. 9). Here four major generations (I – IV) show a dome-like, but slightly asymmetrical uplift. In the north, the four generations show a simple tilt to the east with decreasing dip angles for the lower generations. The Muddus plains form the flank of the Northern Scandes, while a hilly relief forms the flank of the Southern Scandes. This has been explained as a Mesozoic etch-surface protected by a Cretaceous cover until late Neogene (Lidmar-Bergström 1995). This provides an argument for a late uplift of the Southern Scandes (Lidmar-Bergström 1999) and a correlation between the lowest surface of the palaeic relief in the south with the Muddus plains in the north (Lidmar-Bergström & Näslund 2002). A correlation of surfaces identified by

| Table 1: Correlation of erosion surfaces between the Northern and Southern Scandes |
|-----------------------------------|---------------|---------------|---------------|
| **Southern Scandes**              | **Northern Scandes E** | **Northern Scandes W** |
| Lidmar-Bergström et al. 2000      | Wråk 1908      | Wråk 1908      | Wråk 1908      |
| I > 2000                          | 0 1850 - 1500  | Likka         |
| II 1800 - 1500                    | 1 1750 - 1100  | Likka         |
| III 1400 - 1200                   | 2 1500 - 800   | Tuipal        |
| IV 1200 - 800                     | 3 1200 - 650   | Borsu         |
| V 700 – 500 valley gen.           | 4 800 - 400    | Muddus        |
| Valley generation                 | MP 550 – 300   | 4a 900 - 800  |
| The strandflat                    | Eastern flank  | 4b 700 - 500  | Guris         |
|                                  |                | Pakko         | Geba          |
| MP = Muddus Plains                |                | Lule          | Raisa         |

The strandflat
The causes of uplift are not clear (Olesen et al. 2002). Geophysical investigations (Ebbing & Olesen 2005) show a flexural rigidity for the lithosphere of the Southern Scandes, while no significant rigidity can be established for the Northern Scandes. Furthermore, the isostatic gravity and geoid residual indicate isostatic support by a flexural rigidity for the lithosphere of the Southern Scandes, while no significant rigidity can be established for the Northern Scandes. The isostatic residual supports the view that there is a difference in style and timing of uplift between the Northern and Southern Scandes. The late uplift amounts to only 500 m in the Northern Scandes, compared to about 1000 m in the Southern Scandes (Lidmar-Bergström & Näslund 2002). The hinge line (Fig. 7) marks the border between the hilly relief (sub-Cretaceous re-exposed surface) and the Muddus plains east of the mountains (cf. Lidmar-Bergström & Näslund 2002). Here it also coincides with a change in Bouguer gravity data (Olesen et al. 2002). The area northwest of the hinge line is a key area for understanding the relation between the Northern and Southern Scandes. The Precambrian basement around the hinge line has several fracture zones (Lidmar-Bergström 1996 figs. 12, 13), which may have been reactivated in connection with late uplift events. The hilly relief southeast of the hinge line ends along the High Coast (Höga kusten) at the Bothnian Sea with several river gaps and river deflections and it is bordered by offshore faults (Lidmar-Bergström 1996, Fig.17).

Remnants of four major erosion generations (surface or valley generations) are preserved within the Northern Scandes. The landscape generations are all inclined to the ESE, are 300 – 400 m apart within the mountains, and merge more or less east of the mountains. Generations 1 and 2 form the low summits and generation 3 and 4 merge and form the floor of the Muddus plains. The four landscape generations indicate four discrete Cenozoic uplift events and a pronounced east-west asymmetric uplift style different from the Southern Scandes. The asymmetrical uplift is thought to have been guided by reactivation of fault systems along the Norwegian coast.

Correlation of level IV in the Southern Scandes with level 4/4b and the Muddus plains in the Northern Scandes shows a difference in amount of the latest major uplift of about 500 m. A hinge line is defined between the two uplift centres.

The reconstruction of a preglacial reference surface makes it possible to estimate the amount of glacial/fluvial erosion since the beginning of the Late Cenozoic glaciations. This amounts to between 200 and 400 m within several valleys with maximum values in valleys that hosted the major outlet glaciers. Glacial erosion by the major outlet glaciers has also been effective over larger areas at the mountain front with erosion values up to 250 m. Glacial erosion by westward moving ice has had a major effect along the two main valleys.

Preservation of surfaces depends greatly on bedrock resistance and the uppermost surfaces are only preserved within the Seve nappes. Glacial erosion has been selective and has not destroyed the sequence of stepped surfaces.

The large-scale relief of the Northern Scandes and their eastern flank mainly reflects a Cenozoic fluvial landform development in response to Cenozoic uplift events. The difference in shape between the Northern and Southern Scandes is interpreted to reflect differences in style and timing of uplift and is therefore important for understanding the causes of uplift.

Landform analysis of large landscape forms using digital topographical data is useful for conclusions concerning major processes responsible for shaping of landscape.

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