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Review of denudation processes and quantification of weathering and erosion rates at a 0.1 to 1 Ma time scale

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Preface

This document contains information on surface weathering and erosion in the Forsmark and Laxemar areas to be used in the safety assessment SR-Site. The report was written by Mats Olvmo, University of Gothenburg.

Stockholm, June 2010

Jens-Ove Näslund

Person in charge of the SKB climate programme

Contents

1 Introduction

The Swedish Nuclear Fuel and Waste Management Company (SKB) is responsible for the management of spent nuclear fuel and radioactive waste generated within the Swedish nuclear power program. SKB plan to submit an application to build a deep geological repository for spent nuclear fuel at the Forsmark site. The Laxemar site was included in the present study as part of the localization process. An important part of the application is the assessment of long-term repository safety.

The deep geological repository shall keep radiotoxic material separated from man and environment for 100,000 years and more. Over the last 800,000 years or so about 100,000 year long glacial-interglacial cycles have dominated climate variation. The time span of a glacial-interglacial cycle, approx. 100,000 years, is similar to the time it takes for the radioactivity in the spent fuel to decay to levels comparable to the activity in the natural uranium that was once used to manufacture the fuel. The 100,000 year time frame is thus important for analysis of long-term safety. In addition, SKB need to discuss safety issues of the repository over a 1 Ma time perspective. In this context it is necessary to discuss the effects of erosion, weathering and uplift for the Forsmark and Laxemar regions for the 100 ka and 1 Ma time perspectives.

The main purpose of this report is to provide information on denudation processes for Forsmark and Laxemar by evaluating the effect of long term landform development in these regions. One issue is to investigate to what extent landform development in the future could reduce the thickness of the geological barrier within the 100 ka and 1 Ma time frames. In this context the mechanisms of denudation are of main interest. The report includes a short introduction to the concept of landform development, a review of different denudation processes that are of importance in the two regions, a review of denudation rates in different geological contexts, a brief description of the long-term landform development in South Sweden and finally a description and evaluation of the Forsmark and Laxemar regions.

2 Methods

The first part of this report which focuses on landform development, denudation processes and rates as well as the long-term landform development in South Sweden is based on a literature study. The second part which deals with the two areas of concern is based on map analysis and observations from short field trips. The maps presented in the report are produced in order to describe different aspects of the relief of the study areas and to put them into a regional perspective. All maps are based on elevation data (DEM) delivered by the Swedish Land Survey (Lantmäteriet) with a spatial resolution of 50 m/pixel. This resolution is considered good enough to describe the relief at a regional scale. The digital elevation data is processed in different ways by using the ESRI, ArcMap 9.3 software. The maps of relative relief was obtained by using neighbourhood statistics, setting the neighbourhood to 6 map units (300 m) and calculate the maximum elevation range within that area. The hillshade maps were constructed by illuminating the DEM from different directions and with different illumination altitudes in order to get the best expression of the relief. Field work was performed in four days in each area. The field work was done in order to get a general overview of the study areas and to document (photography) some characteristic geomorphological features of the areas.

3 Landform development and denudation processes

3.1 Concepts and models

Landform development through time is complex cf. /Thornes and Brunsden 1977/. Landforms change as a result of tectonic plate movements and climate change and their effects on denudation processes. In general, no landform exists for ever. They are created, develop and disappear and are replaced by others. The landforms and landform development on Earth and other planets are studied within the science of geomorphology. The history of geomorphological science presents different approaches to the study of landforms. A comprehensive text on concepts and theories in geomorphology is given by /Thorn 1988/.

Early models of landform development were often qualitative, based on few measurements and focused on the development of landscapes levelled down to areas of very low relief labelled peneplains. It is interesting to note that early geomorphologists /cf. Ramsay 1863, Powell 1876/ paid so much attention to such *planation* surfaces. The recognition that many Precambrian land masses were levelled down to plains that expose crystalline rocks formed at considerable depths in the crust has been and still is a challenge for geoscientists.

Peneplanation was also included in the early cycle theory of /Davis 1899/, which had great impact on geomorphology for half a century. The theory presented by /Davis 1899/ was an attempt to describe the interaction between river incision and crustal uplift of crustal blocks that lacked initial relief. His model explains the development of relief through different stages of maturity ending up with a nearly flat surface with some residual relief, the peneplain.

The cycle theory of Davis was a general conceptual model of landform development lacking a quantitative approach. /Gilbert 1877/ who was contemporary with Davis was the first to introduce a process-response concept in geomorphology. /Gilbert 1877/ introduced the term dynamic equilibrium to refer to any change in a geomorphic system that causes the process or processes to operate in a way that tends to minimize the effect of change; a negative feedback. According to this idea the system adjusts over time so that process rates change in order to minimize changes within the system. At that time Gilbert´s approach became unrecognised, probably as a consequence of the popularity of Davis cycle theory.

As a consequence of what is known as the quantitative revolution in geomorphology of the 1950s and 1960s, the use of the concept of equilibrium came to the fore. /Hack 1960/. The concept of /Schumm and Lichty 1965/ of the gradual episodic directional change has possibly become the dominant regulatory principle governing landscape development over geologic time scales. This view is a modification of the traditional monotonic decline in landscape elevation envisioned by Davis /Rhoads and Thorn 1993/.

/Büdel 1957, 1982/ made a significant contribution to the understanding of landform development by emphasizing the importance of two types of surfaces of geomorphic activity. According to his idea, sub-surface etching (deep weathering) at the weathering front and surface erosion act together to lower the landsurface. The idea of Büdel has become the basis of many geomorphological explanations of landform development at a continental scale since weathering mantles and remnants of weathering mantles are common features in different setting, e.g. /Ollier 1988, Thomas 1989a, b, Twidale 1990, Lidmar-Bergström 1996, Lidmar-Bergström et al. 1997, Olvmo et al. 2005/.

The long term development of landforms is, however, complex and many landforms in Precambrian shield areas have a long history spanning periods of exposure, burial and re-exposure. This means that many landforms in these areas are relict features that are re-exposed from below a former cover of sedimentary rocks and thus not related to processes operating at present. This concept is especially true in Sweden where it has been known for long that re-exposed sub-Cambrian and sub-Mesozoic landforms cover extensive areas cf. /Högbom and Ahlström 1924, Rudberg 1954, Lidmar-Bergström 1988, 1991, 1996/.

3.2 Denudation

Denudation is defined as the overall degradation and levelling of continental land masses cf. /Ahnert 1996, Smithson et al. 2008/. Denudation is achieved by different exogenic processes, including weathering, mass wasting and erosion by wind, running water, waves and glaciers. The energy needed for the denudation processes is gained from endogenic and exogenic sources.

3.3 Weathering –the first step in the process of denudation

Weathering exerts the most fundamental control on denudation and is the driver of, or limiting factor, in landscape evolution /Turkington et al. 2005/. Several authors have shown the significance of differential weathering in landscape evolution cf. /Ollier 1960, Thomas 1966, 1994/. Deep weathering has been considered important in humid tropical regions for long, however, the fundamental role of deep weathering in different environmental settings also outside the tropics has recently been pointed at /Migoń and Thomas 2002/.

Weathering can be defined as structural and/or mineralogical breakdown of rock and soil materials by physical, chemical and biological processes at or near the surface of the Earth e.g. /Reiche 1950, Keller 1957, Ollier 1969, Selby 1993, Whalley and Warke 2005/. The definition indicates that weathering occurs when minerals/rocks are exposed to temperatures, pressures and moisture conditions characteristic of the atmosphere and hydrosphere, that is in an environment that differs significantly from the conditions in which most igneous and metamorphic rocks, as well as lithified sedimentary rocks were formed. Therefore, the alteration of rocks by weathering forms new materials (minerals) that are in equilibrium with conditions at or near the Earth's surface.

By definition weathering occurs *in situ* and does not directly involve erosion. This means that it leads to the formation of a residual material that differs from the parent, unweathered rock with respect to its physical and chemical properties. Weathering normally lowers the strength of rock and increase permeability of the surface material and thus makes it more prone to mass wasting and easy to erode by running water, glacier, wind etc. In addition, it is also an important prerequisite for the widespread development of flora and fauna on land by releasing nutrients for plants and other organisms.

3.3.1 Weathering processes

Weathering is generally divided into physical, chemical and biological components. Physical or mechanical weathering occurs when volumetric expansion and related alteration of stresses lead to failure and disintegration of the rock. For example, volume changes due to decreased overburden and stresses can result in the creation of fractures at various scales. Crystallisation and volumetric alteration of salt crystals, freezing of water and freeze-thaw effects as well as thermal fatigue due to repeated (diurnal) heating and cooling as well as thermal shock associated with fires, may also cause physical weathering.

Chemical weathering comprises reactions between rock minerals and water. Examples are solution of minerals, carbonation, hydrolysis, hydration, and oxidation and reduction. Common to chemical weathering processes is that they depend on water composition, for example pH, salinity, CO_2 and redox potential. The prevailing temperature is another important parameter determining the type and efficiency of chemical weathering. Biological weathering comprises biochemical alterations of rock minerals and mechanical impact caused by drying of lichen and fungi, boring into rock by biota and root penetration into fractures and joints.

3.3.2 Deep weathering

The term deep weathering is normally used to describe the process by which a more or less thick mantle of altered rock is formed by in situ weathering cf. /Ollier 1969/. Alteration of rock by deep weathering occurs to depths of tens or even a hundred meters. Deep weathering may be the result of a progressively falling water table and downward extension of the oxidation zone, but weathering may occur below a water table in reducing conditions /Ollier 1988/. Hydrolysis is the dominant process below the water table and silicate minerals react with water to form metallic ions and hydroxyl ions in solution leaving a residuum of clay. The production of hydroxyl ions enhances the breakdown of silicates since the pore fluid becomes more alkaline. A consequence of this change in chemistry is diffusion transport of ferrous iron towards the oxidizing zone where it is precipitated and converted to ferric hydroxide. In this way layers of iron enriched ferricretes, 1 m thick may be formed in 10,000 years /Selby 1993/.

Deep weathering is often referred to as tropical weathering and has often been used as an indicator of former humid tropical climate conditions. However, this is not always appropriate although the process may be of major importance in these environments. As suggested by /Migoń and Lidmar-Bergström 2001, 2002/ it is likely that deep weathering is time and space continuous, although it operates with different intensities and the balance between rates of saprolite production and surface erosion have shifted through time.

The association of deep weathering and humid tropical climates may have been biased by the great focus on clayey weathering mantles dominated by kaolin minerals. As /Migoń and Lidmar-Bergström 2002/ put it "*These kaolinisation periods coincide with periods of humid climate, while during arid phases other clay minerals were preferentially formed such as chlorites, smectites, and mixed layer minerals.*" Many weathering mantles in temperate regions are poor in clays and differ substantially from deep kaolinised types. These saprolites are often referred to as grus. According to /Migoń and Thomas 2002/, grus is an ill-defined product of deep weathering of coarse-grained rocks whose relationships to other weathering changes remain unclear. /Migoń and Thomas 2002/ conclude that grus weathering is connected with weakening of the rock fabric by development of microcracks, biotite expansion, and initial alteration of plagioclase and may originate either beneath the surface or at greater depths within a weathering profile, see also /Lidmar-Bergström et al. 1997/.

Grus mantles are widespread across climatic zones and in contrast to kaolinisation climate seems to play less important control on grus weathering. Instead, topographic and also petrographic factors appear to play key roles in the development of grus. This may be regarded as a response of weathering systems to rapid relief differentiation which may explain the association between grus mantles and areas with moderate to high relief /Migoń and Thomas 2002/.

3.3.3 Factors affecting weathering

In nature weathering is a complicated process and many different factors affect the dominant type of process as well as the weathering rate. However, as weathering is the result of the exposure of rocks and minerals formed at different depths to atmospheric conditions at the surface, two factors may be considered most important, i.e. climate and rock composition. The function of weathering is to reach balance between the external forces (climate) and material properties (rock composition). However, as these factors are constantly changing it can be questioned if true equilibrium can be established at the surface e.g. /Bland and Rolls 1998/.

Climate

Climate controls both the type of process involved in weathering as well the weathering rate. Temperature is a climatic factor that affects the reaction rate. Energy is needed in order for a reaction to occur. If enough energy is put into a reaction chemical bonds can break and new products can be formed. A rise in temperature may increase the amount of energy in a reaction and hence affect the rate. The relationship between temperature and the chemical reaction rate in general is described by the Arrhenius equation:

$$k = \mathrm{A}e^{-\mathrm{Ea/RT}} \quad (3\text{-}1)$$

where k is the rate constant, A is the frequency factor, E is the activation energy, R is the gas constant and T is the temperature. This means that with a rise in temperature by 10°C the reaction rate rises by a factor of two /Bland and Rolls 1998/. As a consequence a considerable variation in the efficacy of weathering environments exists on Earth. /Thomas 1994/ suggests that the temperature factor increase the weathering reactions by a factor of four when comparing low latitude and high latitude climates. /Bland and Rolls 1998/ argue that the weathering reaction rate increase by a factor of ten between polar and tropical regions.

Hydrology

Water plays an important role in most weathering processes. Chemical reactions need water to occur, but water is also crucial to many mechanical weathering processes. The role of water is, however, more difficult to quantify compared to temperature. In chemical weathering the most important function of water circulation within the rock mass is to prevent equilibrium in weathering solutions to be attained e.g. /Thomas 1994/. Precipitation is therefore important and there is certainly a contrast between different climate zones at present, both considering the amount of precipitation and its distribution in time. For weathering to take place effectively it is important that water can infiltrate and circulate through rock and/or soil. Bare solid rock surfaces are normally relatively immune against weathering since the water will run off or evaporate quite rapidly.

Characteristics of the parent material (the rock)

Besides the external factors discussed above, different properties, such as mineral and chemical composition of the parent material are important for the rates of weathering. Different mineral composition might be expected to respond different to weathering processes. A common method to classify minerals in order of their stability is to arrange them by the order of frequency of occurrence in sedimentary rocks of increasing age cf. /Pettijohn 1941/.

Fractures and other partings play an important role. /Selby 1993 p. 127/ suggests that the formation of joints itself is the most important weathering process, even though their formation seldom is looked upon as a weathering process. However, joints are very important for other weathering process, mechanical, chemical and biological, not at least because of the effect on water circulation in rocks /Twidale and Campbell 1995/.

3.3.4 Weathering rates

The definition and especially the measurement of weathering rate are not straightforward. As /Thomas 1994/ puts it "*it might be thought that this* (the measurement of weathering rate) *was the topic capable of precise numerical expression*", but finds that this seldom is the case. The weathering rate could be looked upon as a measure of loss of material per unit area over a certain amount of time. In many cases the erosional transfer of sediments and solutes are therefore used as an approximation for the weathering rate. /Phillips 2005/ defines weathering rate as the rate at which parent material is converted to weathering products and residuals.

Since saprolite formation is usually an isovolumetric transformation of the bedrock it means that erosion and resulting lowering of the topographic surface do not necessarily accompany rock weathering at depth. /Thomas 1994/ suggests that weathering rates can be considered using two different approaches, namely the rate of saprolite formation or the rate of mineral transformation, but notes that these two approaches are not necessarily comparable. The problem becomes even more complicated when different time scales are involved. It is for instance not a simple task to extrapolate data on present day weathering rates to rates over geological time spans.

Present day weathering rates are commonly calculated by geochemical mass balance studies of watersheds cf. /Pavich 1985, 1989, Velbel 1985, 1986/. A compilation of results from that kind of evaluations are presented by /Thomas 1994/ and in Table 5-1 in the present report. The results are given as a measure of the surface lowering per unit time and are thus a measure of denudation rate rather than weathering rate. The results range from 2–48 m/Ma and apply for crystalline igneous and metamorphic rocks in the United States. The means of the data presented by /Thomas 1994/ using different methods are 20.8 m /Ma (mass balance) and 22.5 m/Ma (rates of alteration), respectively. The extremes are 2 m/Ma (min) and 48 m/Ma (max) for the mass balance calculations and 2 m/Ma to 50 m/Ma for alteration rate calculations. This implies that the range of weathering rate in crystalline rocks and in different climatic settings ranges between 2 and 50 m/Ma.

The development of new techniques including fission-track thermochronology /cf. Vernon et al. 2008/ and cosmogenic radionuclides has increased the opportunities to calculate weathering rates. Cosmogenic radionuclides have been widely used for surface exposure dating in different settings cf. /Burbank et al. 1996, Owen et al. 2002, Stroeven et al. 2002, Zehfuss et al. 2001/. /Granger and Riebe 2007/ and /Kirchner et al. 2006/ have shown that cosmogenic nuclides such as ^{10}Be and ^{26}Al can be used to measure rates of physical erosion and chemical weathering over millennial to 10,000-year time scales. The results of /Kirchner et al. 2006/ imply that the strength of climate change feedbacks between temperature and silicate weathering rates may be weaker than previously thought. They also found that chemical weathering rates may be limited by the rates at which fresh minerals are supplied to soils by erosion, implying that tectonic uplift may be an important regulator of long-term chemical weathering rates in mountainous, granitic landscapes such as Sierra Nevada in California.

4 Erosion processes

Erosion can be defined as the removal and transport of bedrock and earth materials by a moving natural agent, such as air, water or ice. Erosion is often preceded by weathering and followed by sedimentation. In this section a brief description is given of the main mechanisms of fluvial and glacial erosion.

4.1 Fluvial erosion

Less than 0.005% of the global water is stored in rivers nevertheless they are one of the most, if not the most, potent erosional forces operating on the Earth's surface cf. /Knighton 1998/. Rivers cut valleys, transport sediments and deposit their loads in a variety of depositional environments, such as flood plains and deltas. Vertical erosion by rivers is a striking feature of the world's mountainous areas where deep valleys dissect the landscape and form steep slopes, thereby creating the conditions for mass movement processes that are closely linked to fluvial vertical erosion.

The morphology of rivers and the processes of erosion and deposition depend on the interaction between the fluid flow in the channel and the properties of the materials in the channel boundary. Basically the erosion and transport by flowing water is a function of the kinetic energy:

$$E_k = mV^2/2 \qquad (4\text{-}1)$$

where m is the mass of the water and V is the flow velocity /Ahnert 1996/.

The Chezy equation is also valid:

$$V = CD^{0.5} S^{0.5} \qquad (4\text{-}2)$$

where C is the Chezy coefficient, D is the water depth and S is the topographic gradient. The substitution of equation (4-2) for V in equation (4-1) shows that that the kinetic energy of flowing water is directly proportional to the product of depth and gradient. Hence, the depth-slope product is the basic controlling parameter for stream erosion, and implies that mountainous streams of high discharge have a high erosion competence.

The relationship between some of the controlling variables in the fluvial system is conceptualized by /Bull 1979/. The relationship is expressed as a ratio describing the threshold of critical power:

$$\text{stream power (w) /critical power} = 1 \qquad (4\text{-}3)$$

in which

$$w = \gamma QS/\text{width} = \gamma dsu = Tu \qquad (4\text{-}4)$$

where γ is the specific weight of water, Q is the discharge, S is the gradient, d is the main depth of flow, u is the mean flow velocity and T is the mean boundary shear stress. The critical power is the power needed to transport sediment load and reflects channel width, depth and flow velocity. This means that in stream reaches where the ratio is higher than one, the stream bed is eroded by vertical erosion, in reaches where the ratio is one lateral erosion occurs, while in reaches with a ratio less than one deposition occurs. In general the most effective vertical erosion by rivers therefore is in low order mountainous steams with steep gradients.

The components of the fluvial morphological process response system according to /Ahnert 1996/ are presented in Figure 4-1. The system is driven by endogenic and exogenic energy supplies and contains three categories of ensystemic (within the system) components; 1) form components, 2) process components, 3) material components.

River incision into bedrock creates dissected topography of uplifted regions of the world. In an attempt to understand the process of channel erosion into bedrock /Seidl and Dietrich 1992/ quantified a simple erosion law by measuring drainage areas and slopes on both principal channels and

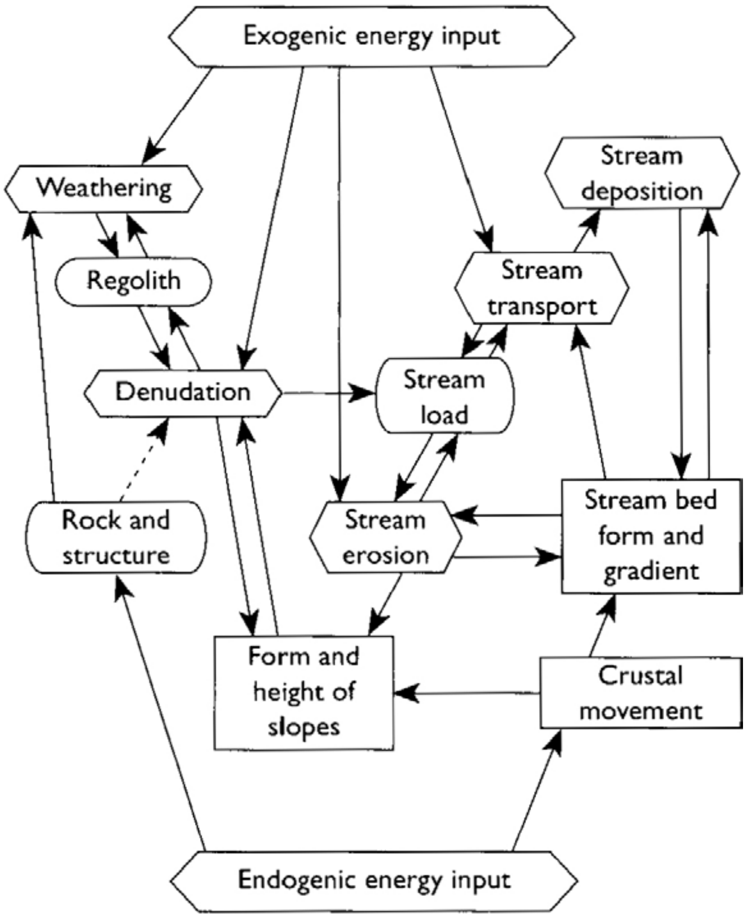

Figure 4-1. *Components of the fluvial morphological process response system according to /Ahnert 1996/.*

tributaries. Their data suggest that both a bedrock-cut tributary and main stream will incise at the same rate at their confluence if the ratio of main stream to tributary drainage area equals the ratio of tributary to main stream channel slope at the junction. This suggests that the erosion across several tributary junctions is linearly related to stream power. They also found that tributary slopes greater than about 0.2 deviate from this linear prediction, probably because debris flows scour these steep tributaries.

/Seidl and Dietrich 1992/ also found that the common elevation of tributary and main stem may result from the upslope propagation of locally steep reaches generated at tributary mouths. This propagation continues only to the point on the channel where the channel slope is too steep to preserve the over-steepened reach, or knickpoint, and debris flow scour dominates channel erosion. They conclude that there are three general mechanisms by which bedrock channels erode: (1) vertical wearing of the channel bed due to stream flow, by processes such as abrasion by transported particles and dissolution; (2) scour by periodic debris flows; and (3) knickpoint propagation. This suggests that river incision is a complicated process involving erosion both by "normal" stream flow and mass movement processes. It should be noted that these mechanisms occur preferably in steep mountainous terrain. In many upland or lowlands, such as southern and central Sweden, or even mountains of medium height such as most Central European ones these processes probably are not at all important.

4.2 Glacial erosion

4.2.1 Processes

Glacial erosion is traditionally divided into two components; plucking and abrasion. Plucking is the process by which rock fragment of different size is loosened, entrained and transported away from the glacier bed, while abrasion is the grinding of the substrate by rock fragments held in the moving glacier bed cf. /Sugden and John 1976, Benn and Evans 1998/. The former is responsible for the characteristic irregular and fractured surfaces of lee sides of bedrock bumps and small hills in formerly glaciated areas.

Failure leading to loosening of fragments is known to be caused by stresses set up by differential ice load and high water pressures at the ice-bedrock interface cf. /Sugden and John 1976, Drewry 1986, Benn and Evans 1998/. Many minor fracture features, such as chattermarks and crescentic gouges, observed on glacially affected rock surfaces, indicate that normal stress at the glacier bed may be sufficient to cause failure in some rocks. However, the role of pre-existing weakness such as joints, cracks and foliation, and also pre-glacial weathering should not be underestimated cf. /Olvmo and Johansson 2002/.

Abrasion is the process whereby the bedrock beneath a glacier is scoured by debris carried in the basal layers of the glacier cf. /Sugden and John 1976, Drewry 1986/. The process leads to striation and polishing of bedrock surfaces cf. /Benn and Evans 1998/ and is typical of the stoss side of rock bumps in formerly glaciated terrain. Many factors control the effectiveness of glacial abrasion cf. /Benn and Evans 1998/. The relative hardness of the overriding clast and the bedrock is important and the process is most effective when the overriding clast is much harder than the substratum. The force pressing the clast against the bed is decisive and depends in turn on the shape of clast as well as the motion of the clast during transport. Other factors related to the availability of clasts in the basal ice layers, such as clast concentration, removal of debris and availability of basal debris are also of major importance. Water flowing in subglacial channels between the ice and the bed erodes the surface in the same way as water on an ice-free bed.

4.2.2 Magnitude of glacial erosion

Glacial erosion is complex and depends on material properties of ice and rocks, glacier dynamics, friction and lubrication and sub-glacial hydrology. Therefore the magnitude of glacial erosion differs widely both in time and space. At a continental scale, the large scale pattern of glacial erosion probably is controlled by ice sheet thermal regime and topography of the subglacial landscape. Based on a simple glaciological model /Sugden 1977, 1978/ made a reconstruction of the thermal regime of the Laurentide ice sheet. The reconstructed thermal pattern shows an inner wet-based area and an outer cold-bed area, which broadly corresponds to the pattern of glacial erosion indicated by the distribution of erosional landforms. The most intense erosion as indicated by areas with high lake density coincide with the transition zone between wet-based and cold based ice in the model, which probably favors plucking and debris entrainment.

/Näslund et al. 2003/ used a numerical ice sheet model to study regional ice flow directions and glacial erosion of the Weichselian ice sheet in Fennoscandia. In their study, ice sheet model results from different time periods during the Weichselian were extracted for five regions and compared with information on flow directions obtained from current conceptual geological models based on field data. The similarities between the computer model and the conceptual model were strikingly good both with respect to ice flow and timing. They also introduced a new quantity, basal sliding distance, describing the accumulated length of ice that has passed over the landscape by basal sliding, and suggested that this entity could be used as a proxy for glacial erosion. The results indicate high basal sliding distance values in SW Sweden/SE Norway, in Skagerrak, and along the Gulf of Bothnia, implying relatively large amounts of glacial erosion in these regions. On elevated parts of the Scandinavian mountain range and on adjacent plains in the east the basal sliding distance values are low, implying weaker glacial erosion, which is fairly in agreement with geological and geomorphological evidence cf. /Rudberg 1967, Lagerbäck and Robertsson 1988, Riis 1996, Stroeven et al. 2002, Olvmo et al. 2005/. The method of estimating glacial erosion by simulated basal sliding distance /Näslund et al. 2003/ was further developed by /Steiger et al. 2005/ who introduced a normalization of the sliding values by the duration of ice cover over a particular site. /Steiger et al. 2005/ also set up a relationship between normalized sliding distance and rate of glacial erosion.

Another approach to the issue of glacial erosion is presented by /Hallet et al. 1996/ who made a comprehensive review of glacial erosion rates based on sediment yields. They found that rates of glacial erosion vary by many orders of magnitude from 0.01 mm yr^{-1} for polar glaciers and thin temperate plateau glaciers on crystalline bedrock, to 0.1 mm yr^{-1} for temperate valley glaciers also on resistant crystalline bedrock in Norway, to 1.0 mm yr^{-1} for small temperate glaciers on diverse bedrock in the

Swiss Alps, and to 10–100 mm yr^{-1} for large and fast-moving temperate valley glaciers in the tectonically active ranges of southeast Alaska. These major differences highlight the importance of the glacial basal thermal regime, glacial dynamics and topographic relief on the rates of glacial erosion.

Yet another approach was presented by /Påsse 2004/. In order to estimate the average glacial erosion in the bedrock in non-mountainous parts of Sweden he used seismic data and well data from Sweden and Denmark to calculate the thickness of the minerogenic Quaternary sediments. The average thickness of Quaternary sediments was estimated to be 16 m in the investigated area, which corresponds to 12 m assuming that the whole volume is the result of glacial erosion of fresh bedrock. Since a great part of the sediments likely consist of glacially redistributed Tertiary regolith this figure probably is an overestimation of the glacial erosion depth in the bedrock. Considering this, Påsse (op cit) concludes that the average glacial erosion during a full glacial period may be estimated to between 0.2 m and 4 m. This is in agreement with estimates of glacial erosion in the Precambrian basement based on geomorphological observations /Lidmar-Bergström 1997, Ebert 2009/. Lidmar-Bergström (op cit) distinguishes the estimates of glacial erosion of Tertiary saprolites from glacial erosion of fresh bedrock. The glacial erosion of saprolites is estimated between 10 and 50 m and glacial erosion of fresh bedrock is estimated at some tens of metres /Lidmar-Bergström 1997/.

However, in Fennoscandia as a whole, large spatial differences in thicknesses of Quaternary deposits occur and distinct patterns of glacial scouring and deep linear erosion are observed in places. /Kleman et al. 2008/ point at the relative roles of mountain ice sheets and full-sized Fennoscandian ice sheets for this zonation and use spatio-temporal qualitative modelling of ice sheet extent and migration of erosion and deposition zones through the entire Quaternary to suggest an explanatory model for the current spatial pattern of Quaternary deposits and erosion zones, see Figure 4-2. They use the spatial distribution of fjords and deep non-tectonic lakes for delineating zones of deep glacial erosion, and relict landscapes as markers for frozen-bed conditions. The landscape was classified into a tripartite system of drift thickness on the basis of the amount of exposed bedrock. Areas with "thick drift"

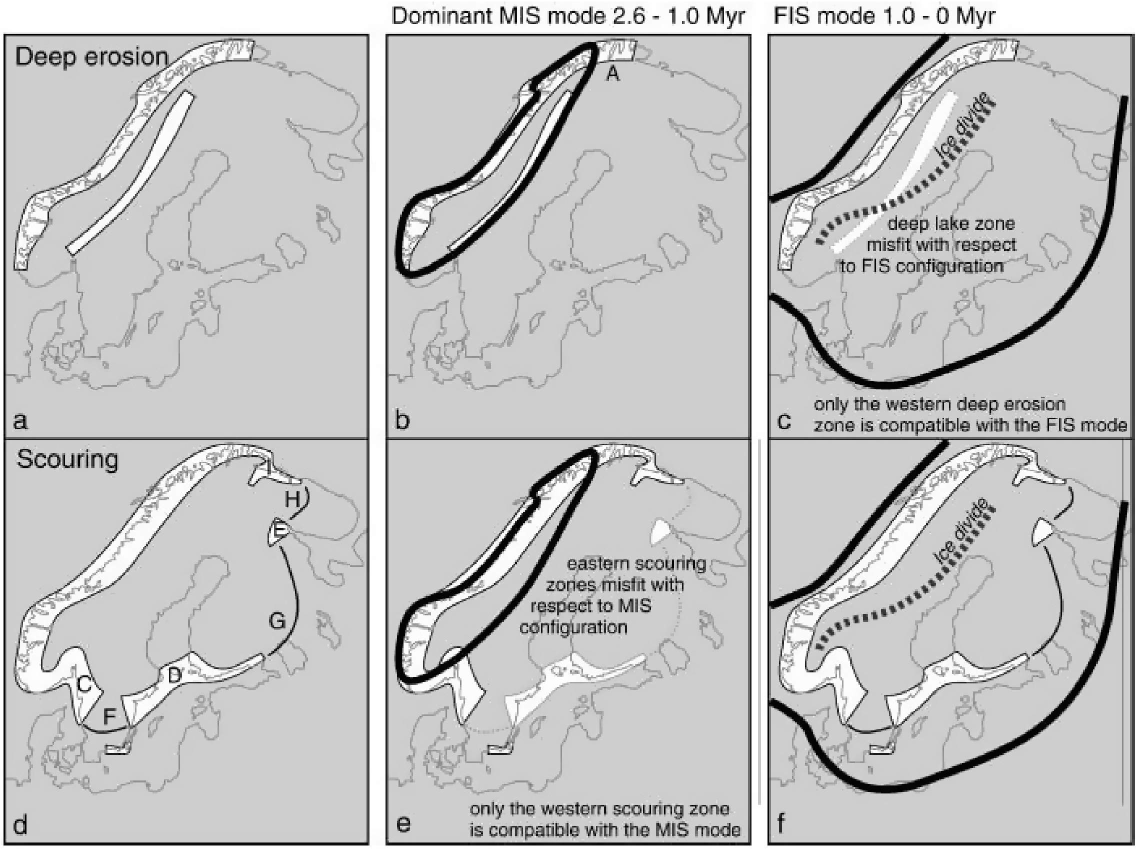

Figure 4-2. Conformance of the zones of deep linear erosion (panels a–c) and scouring (panels d–f) to specific ice sheet configurations and time periods. From /Kleman et al. 2008/.

are defined as areas with no or a few scattered bedrock outcrops, while areas with a dominance of exposed bedrock was classified as "mostly scoured bedrock with patchy drifts". Intermediate fractions of bedrock exposures were classified as "intermediate-thickness drift cover".

A zone of thick drift is found in central and northern Sweden which is suggested to be the combined result of marginal deposition of fluctuating mountain centered ice sheets, during the early and middle Quaternary, and the inefficiency of later east-centered Fennoscandian ice sheets in evacuating this drift from underneath their central low-velocity and possibly frozen-bed areas. A western zone of deep glacial erosion formed underneath both the mountain centered ice sheets and full-sized style ice sheets during the entire Quaternary (Figure 4-2). The eastern zone of deep glacial erosion (east of the Scandinavian mountain range) is exclusively related to mountain style ice sheets, and formed largely during the early and middle Quaternary. The scouring zones formed under conditions of rapid ice flow towards calving margins of full-sized style ice sheets likely reflect process patterns of the last two or three ice sheets.

The three landscape zones differ in their degree of permanence, with the deep erosion zones being a long-lasting legacy in the landscape, more likely to be enhanced than obliterated by subsequent glacial events. The thick drift cover zone, once established, appears to have been surprisingly robust to erosion by subsequent glacial events. The scouring zones appear to be the most recent and ephemeral of the three zones, with possible major alterations during single glacial events /Kleman et al. 2008/.

4.3 Glacial meltwater erosion

Glacial meltwater erosion may be an effective erosion agent both in subglacial and proglacial environments. The sediment concentration of glacial meltwater streams are often high and the flow is often very rapid and turbulent, which mean that flows transitional between debris flows and normal stream flow are common /Benn and Evans 1998/. The erosivity of glacial streams is therefore often high both on bedrock and sediments. Apart from the high erosivity, the mechanisms of glacial meltwater erosion are the same as normal fluvial erosion including abrasion, cavitation, fluid stressing and particle entrainment from cohesionless beds as well as chemical erosion.

Erosion channels formed by meltwater erosion are distinctive features in many glacial environments. Nye channels are subglacially formed ice-directed channels cut in bedrock as well as in consolidated sediments. The channels normally extend from a few tens to few thousands of metres and are up to a few tens of metres wide /Benn and Evans 1998/. Many examples of systems of subglacial meltwater channels are given in the literature. Instructive examples are found in the Cairngorm mountains of Scotland /Sugden and John 1976, Gordon 1993/, the Scandinavian mountain range /Mannerfelt 1945, Gjessing 1960/ and the United States /Booth and Hallet 1993/. Some of the canyon-like channels cut in bedrock mainly in northeastern, central and southeastern Sweden described by /Olvmo 1989/ may be of this type.

Tunnel valleys are known from many areas formerly covered by Pleistocene ice sheets including North America, Germany, Denmark, Poland and the Brittish Islands /Benn and Evans 1998/. They form large, overdeepened channels cut into bedrock or sediments and may extend over 100 km in length and be 4 km wide /Ó Cofaigh 1996/. There is no complete explanation of the origin of tunnel valleys. /Shoemaker 1986/ and /Boulton and Hindmarsh 1987/ explain their formation by steady-state meltwater drainage over subglacial deforming sediments, which tend to creep into the subglacial conduit. This would lead to the formation of a valley that is considerably larger than the conduit. Others suggest that tunnel valleys are formed by catastrophic meltwater floods e.g. /Ehlers 1981, Ehlers and Linke 1989, Wingfield 1990/.

In southeastern Sweden the role of meltwater drainage as a stripping mechanism of old saprolites has been stressed by /Olvmo et al. 1999, 2005/ In these areas meltwater erosion may have been effective at a local scale, but its regional importance is not known. Spectacular meltwater canyons are common in certain parts of Sweden /Olvmo 1989, 1992/ and it has been proposed that at least some of these canyons are weathering etch forms that have been stripped and modified by glacial melt water streams /Olvmo et al. 2005/. In southeastern Sweden it may be possible that glaciofluvial erosion was a major mechanism of stripping of weathering mantles.

5 Long term denudation rates

A compilation of global denudation rates from different geological and topographical setting and are presented in Table 5-1. The rates are based on different methods and therefore not fully comparable. The data show very high rates of denudation in orogenic environment ranging from 1,200–70 m/Myr in the San Bernadino Mountains of California /Binnie et al 2008/ to 200–700 m/Myr in the European Alps /Vernon et al. 2008/. Very low long-term denudation rates are found in the Dry Valleys of Antarctica /Summerfield et al. 1999/ as well as in the shield areas of Fennoscandia and Canada /Stroeven et al. 2002, Ebert 2009, Peulvast et al 2009/. Temperate piedmont areas as exemplified by some Appalachian settings are intermediate between these two extremes /Velbel 1985, 1986, Pavich 1989, Cleaves 1989/.

A convincing attempt to calculate the potential denudation rate in the South Indian shield was made by /Gunnell 1998/. He applied Ahnert´s functional relationship between denudation, relief and uplift to the study area. The functional relationship proposes that the denudation rate can be predicted by the relation D = 0.1535h where *h* is a measure of the local relief (the mean slope). /Gunnell 1998/ found a close similarity between the calculated potential denudation rate and the known long- and short term rates as inferred from apatite fission track data and sediment transport in modern rivers, respectively. The local relief in the area studied by Gunnell varies from 1,400–1,702 m to 0–199 m. The potential denudation within the area varies from 0–19 m/Ma in areas with low local relief to 205–275 m/Ma in areas with high local relief.

Table 5-1. Global rates of denudation in different tectonic and climatic settings.

Area	Setting/geology	Authors	Method	Rate
San Bernadino Mountains, California, USA	Orogenic	/Binnie et al. 2008/	Cosmogenic ^{10}Be Apatite U-Th/He thermochronology	1,200–70 mMyr^{-1}
Brubaker Mts. Penn., USA	Low relief Schist, gneiss	/Price et al. 2008/	Mass balance	4.5–6.5 mMyr^{-1}
Boso Peninsula Japan	Orogenic Sandstones Mudstones	/Matsushi et al. 2006/	Cosmogenic ^{10}Be, ^{26}Al	90–720 mMyr^{-1}
Dry valleys Antarctica	Crystalline	/Summerfield et al. 1999/	Cosmogenic ^{21}Ne	0.26–1.02 mMyr^{-1} 0.133–0.164 mMyr^{-1}
S. Norway	Elev. plain gneiss mica schist	/Nicholson 2008/	Quartz veins, wethering rinds, Schmidt hammer	0.5–2.2 mMyr^{-1}
N. Sweden	Plain crystalline	/Stroeven et al. 2002/	Cosmogenic ^{10}Be, ^{26}Al	1.6 mMyr^{-1}
Rhenish massif Germany	Sedimentary	/Meyer et al. 2008/	Cosmogenic ^{10}Be	4.7–6.5 mMyr^{-1}
Iceland	Basalt	/Geirsdóttir et al. 2007/	Sediment record	5 mMyr^{-1} Holocene average
European Alps	Orogen	/Vernon et al. 2008/	AFT 13.5–2.5 Ma	200–700 mMyr^{-1}
Smokey Mts., USA	Schist, gneiss	/Velbel 1985/	Mass balance	38 mMyr^{-1}
S. Blue ridge	Schist, gneiss	/Velbel 1986/	Mass balance	37 mMyr^{-1}
Pacific NW USA	Orogenic	/Dethier 1986/	Mass balance	33 mMyr^{-1}
Masanutten Ttn. USA	Sandstone, shale	/Afifi and Bricker 1983/	Mass balance	2–10 mMyr^{-1}
S. Piedmont, USA	Piedmont, granit	/Pavich 1986/	Mass balance	4 mMyr^{-1}
S. Piedmont, USA	Piedmont, granit	/Pavich 1989/	Residence time	20 mMyr^{-1}
Baltimore Piedmont, USA	Piedmont	/Cleaves et al. 1970/	Mass balance	4–8 mMyr^{-1}
Baltimore Piedmont, USA	Piedmont	/Cleaves 1989/	Equilibrium model	25–48 mMyr^{-1}
N Sweden	Plain Crystalline rocks	/Ebert 2009/	Palaeosurface reconstruction	1.5–5 mMyr^{-1}
Canada	Plain crystalline rocks	/Peulvast et al. 2009/	Palaeosurface reconstruction	2–8 mMyr^{-1}
India	Escarpment	/Gunnell 1998/	Functional relationship model	205–275 mMyr^{-1}

/Koppes and Montgomery 2009/ address the question whether rivers or glaciers are more effective agents of erosion**.** They present a compilation of erosion rates, which questions the conventional view of glacial erosion as being more effective than fluvial erosion. The compiled data suggest that in regions of rapid tectonic uplift, erosion rates from rivers and glaciers both range from 1 to over 10 mm yr^{-1}, indicating that both are capable of generating erosion rates matching or exceeding the highest rates of rock uplift, see Figure 5-1. When comparing erosion rates over timescales ranging from 10^1 to 10^7 years glacial erosion tends to decrease by one to two orders of magnitude over glacial cycles, whereas fluvial erosion rates show no apparent dependence on time. The conclusion of /Koppes and Montgomery 2009/ is that tectonics control rates of both fluvial and glacial erosion over millennial and longer timescales and that the highest rates of erosion (>10 mm yr^{-1}) generally result from a transient response to disturbance by volcanic eruptions, climate change and modern agriculture. Note the low denudation rates typical of cratons as indicated by the comparatively low rates in Australian rivers, where rates are between 1–10 m /Ma and thus in accordance with the data in Table 5-1.

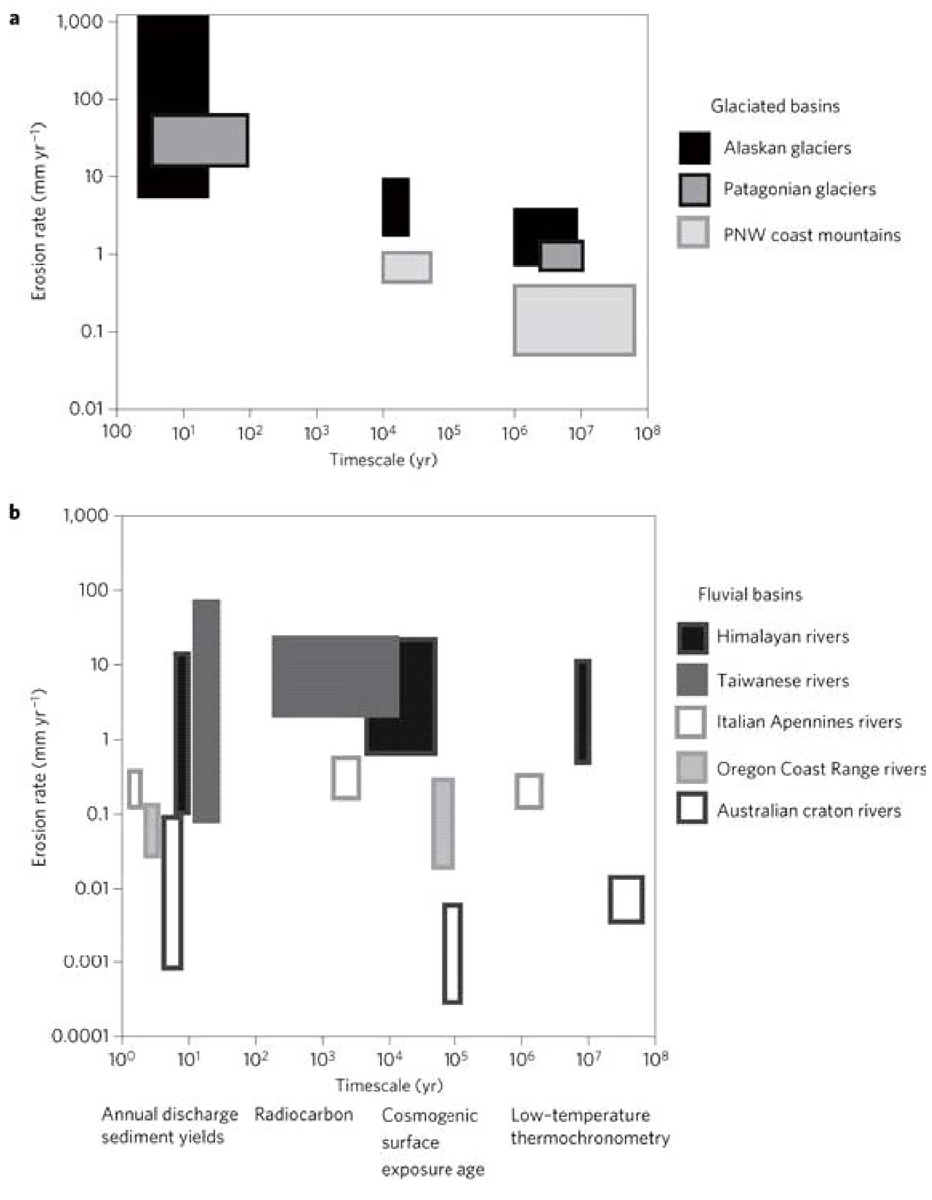

Figure 5-1. *Comparison of short-term and long-term erosion rates from glaciated and fluvial basins. Boxes represent ranges of erosion rates, including errors in estimation (height) and timescale of measurement (width). a, Erosion rates measured from the same or proximal glaciated basins in Alaska, Patagonia and the coast mountains of Washington State in the Pacific Northwest. b, Erosion rates measured from the same or proximal fluvial basins in orogens ranging from most tectonically active to most passive: the central Himalaya, Taiwan thrust belt, Italian Apennines, the Oregon Coast Range and the Australian craton. From /Koppes and Montgomery 2009/.*

6 Long term denudation in southern Sweden

The long term denudation history of southern Sweden is interesting to discuss for two reasons. Firstly, it may be a guide to understand the rate of denudation through time in the Forsmark and Laxemar areas. Secondly, it is important because the different landforms and surfaces that have been recognized may be used as reference surfaces to understand, at least at a regional scale, the magnitude and patterns of glacial erosion which is a main issue with respect to the time perspective discussed in this report.

6.1 The Sub-Jotnian denudation surface

Most of the Precambrian rock of Sweden was formed between 3.0 and 1.4 Ga /Gorbatschev and Gaal 1987/. These basement rocks were deeply eroded already before the Middle Riphean, 1,200 Ma ago, and Jotnian sandstones and shales were deposited on a kind of denudation surface /Lidmar-Bergström 1996/. Today, remnants of the Jotnian sediments are only found in downfaulted positions, for instance in the Bothnian Sea. The sub-Jotnian erosion surface is not significant in the present landscape. It is only preserved in downfaulted basins in central Sweden, for example in the Gävle Basin north of Forsmark. Faulting along the basin margins is probably of Proterozoic age since the sub-Cambrian peneplain cuts across both the basin and the Jotnian sedimentary cover /Lidmar-Bergström 1996/.

6.2 The South Swedish Dome, the sub-Cambrian peneplain and related younger relief

Subsequent mountain building in south western Sweden affected basement rocks during the period 1,200–900 Ma /Gorbatschev and Gaal 1987/. A new period of denudation after 900 Ma levelled out all the differences in relief between south eastern and south western Sweden /Lidmar-Bergström 1996/ as revealed by the very flat topography of the basement just outside the remnants of Cambrian and Upper Vendian sedimentary rocks /Högbom and Ahlström 1924, Rudberg 1954, Lidmar-Bergström 1988, 1991, 1996/.

The most remarkable geomorphological feature of southern Sweden is the low domal structure of the Precambrian basement, labelled the *South Swedish Dome* /Lidmar-Bergström 1988, 1999/. The dome emerges from a cover of Lower Palaeozoic rocks in the east and north and Mesozoic rocks in the south and west (Figure 6-1). The dome reveals two distinct exhumed denudation surfaces. The exhumed sub-Cambrian peneplain forms characteristic relief in large parts in southern Sweden. It extends from below Cambrian cover rocks and is extremely flat close to the cover rocks (Figure 6-2). The peneplain rises from sea level in the east and from the Middle Swedish Lowlands in the north to over 300 m above sea level and ends to the south and west with an erosional scarp.

The relative relief of the sub-Cambrian peneplain is normally less than 20 m /Rudberg 1954, Lidmar-Bergström 1995/ (Figure 6-3), but it is sometimes separated into bedrock blocks that are bounded by low fault-line scarps. The origin of the sub-Cambrian peneplain is not yet fully understood. There are no saprolite remnants preserved on the peneplain, but the basement below Cambrian strata may be kaolinized to a depth of about 5 m /Elvhage and Lidmar-Bergström 1987/. /Lidmar-Bergström et al. 1997/ suggest that the formation of the peneplain was accomplished by etching (that is deep weathering) and subsequent stripping of the weathered material, thus preventing the survival of deep saprolites. They conclude that a prolonged time with stable tectonic conditions, deep weathering, stripping of the weathering mantle and pedimentation produced this extremely flat surface.

In the northeastern part of the South Swedish Dome, sub-Cambrian peneplain is broken in blocks, tilted in different directions /Lidmar-Bergström 1993/, and dissected along structurally controlled lines to form a landscape with joint-aligned valleys /Björnsson 1937, Nordenskjöld 1944/. The relative relief here reaches 100 m and more, but the sub-Cambrian peneplain is still in or close below the hill

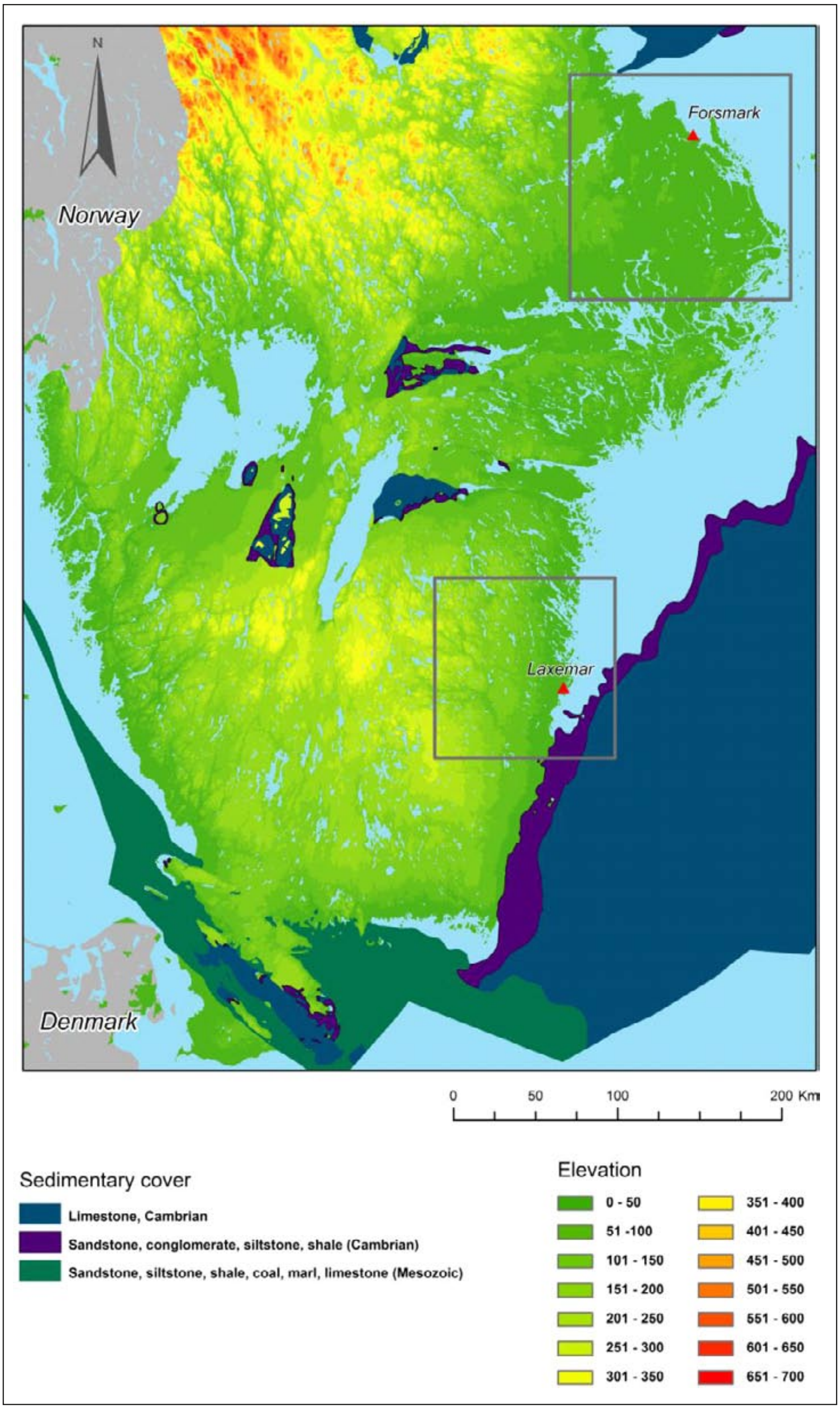

Figure 6-1. *Elevation map of southern Sweden with the extension of Palaeozoic and Mesozoic cover rocks.*

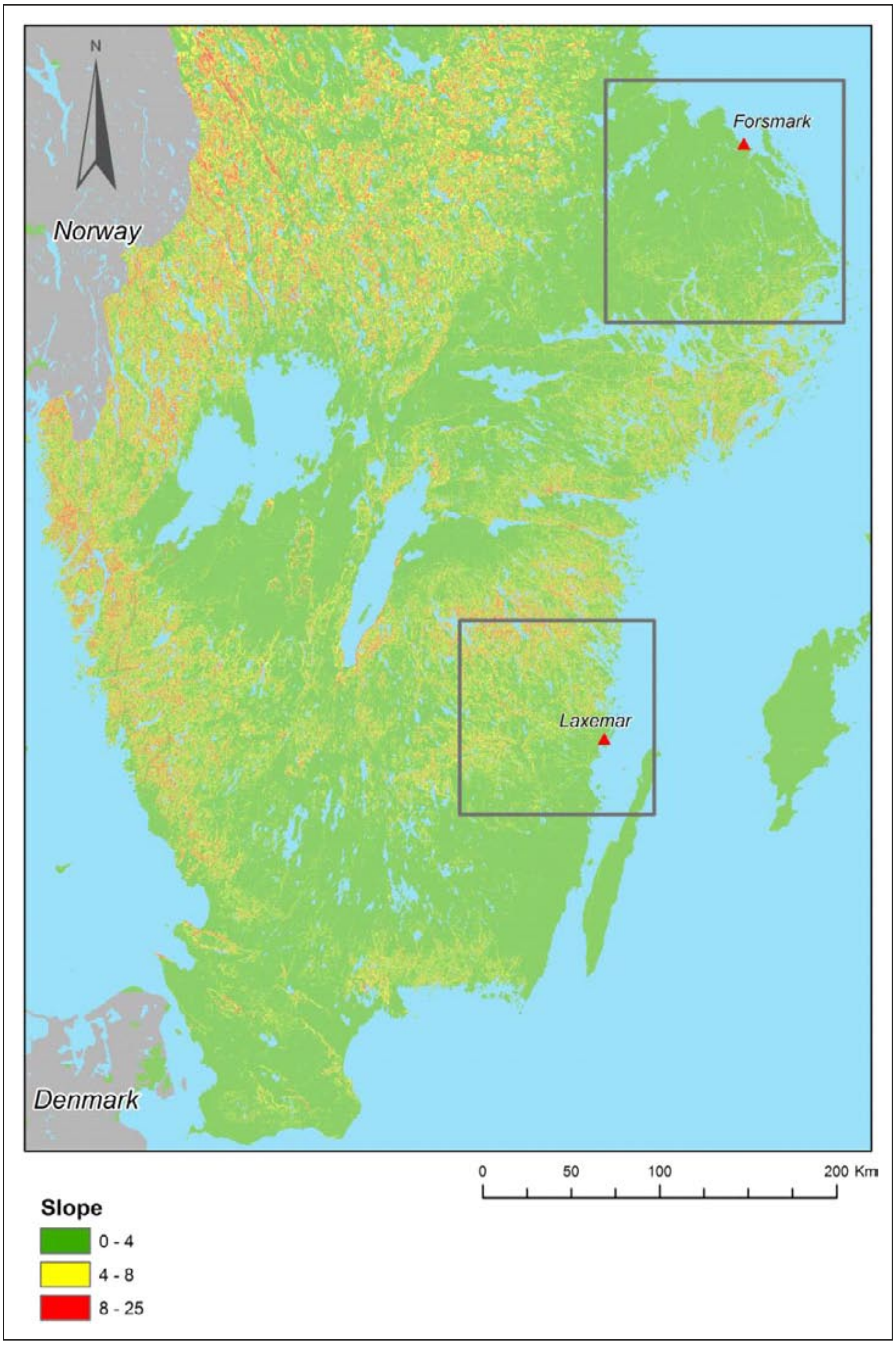

Figure 6-2. *Slope map of southern Sweden. Areas with low slopes show the extension of plains in the Precambrian rocks and on sedimentary strata. The Sub-Cambrian Peneplain forms areas with low slopes close to the cover rocks (compare Figure 6-1). The flat areas in central south Sweden correspond to the South Småland Peneplain.*

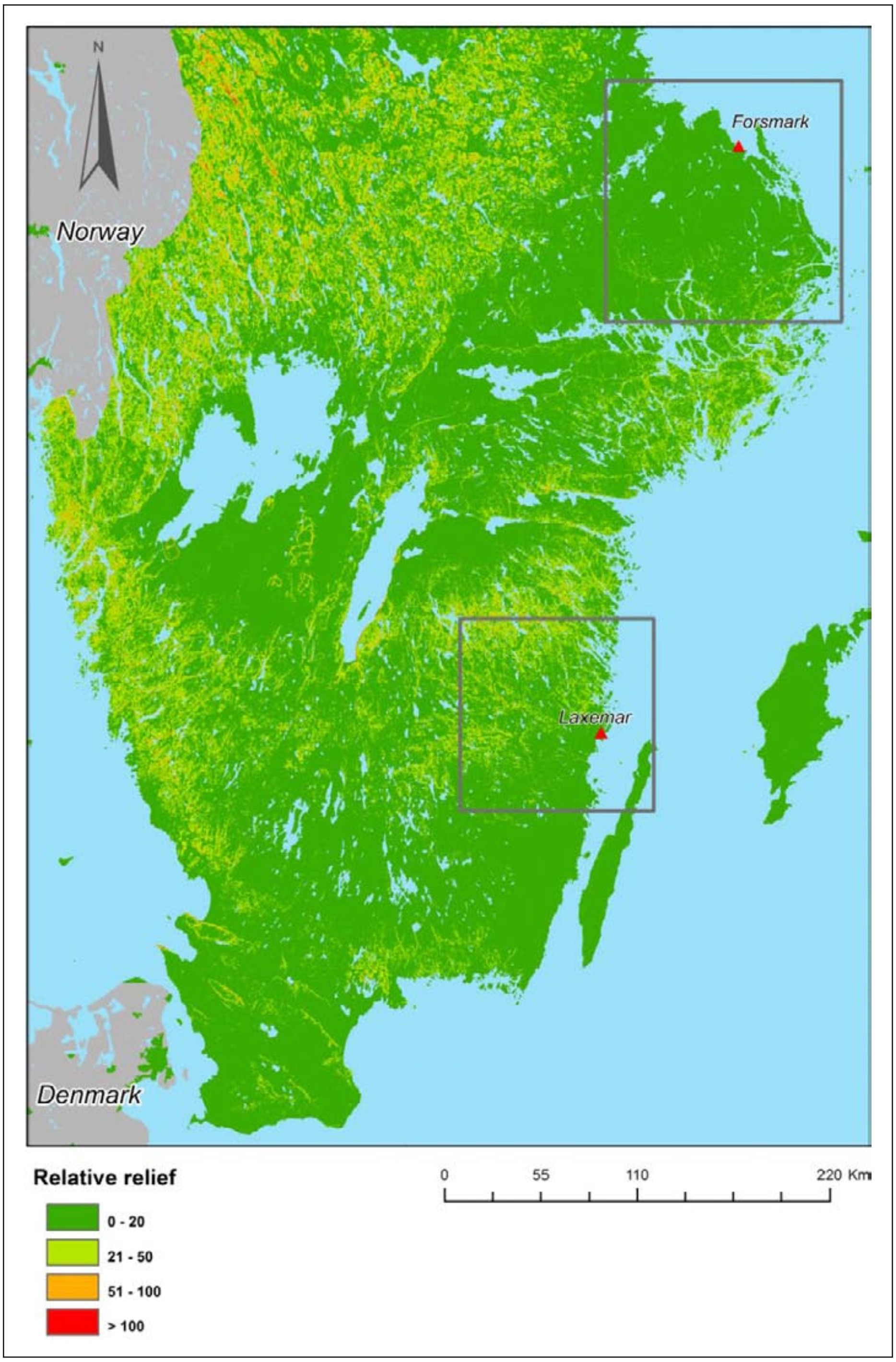

Figure 6-3. *Map of the relative relief in southern Sweden in 4 different classes. Areas with low relief correspond to plains in the Precambrian rocks and on sedimentary strata. Hilly relief is connected with exhumed Mesozoic palaeosurfaces in the south and southwest and in the north. Hilly relief also characterize the uplifted and dissected parts of the Sub-Cambrian Peneplain. Elevation is in m.a.s.l.*

tops (Figure 6-3 and 6-4). In this area two major valleys are incised in the southeast-dipping sub-Cambrian peneplain, while another valley system follows a N–S trending major fracture system. The relief of the dome reflects a diastrophic history since the Early Cambrian /Lidmar-Bergström 1996/. According to /Björnsson 1937/ the fluvial incision of the major valleys predates the Quaternary glaciations. This is in agreement with the recent opinion that the latest phase of uplift, major relief formation, and re-exposure of old surfaces probably occurred in the Neogene /Japsen et al. 2002/.

__Figure 6-4.__ Slope map and height map of southern Sweden showing the structurally controlled landforms of the faulted and dissected Sub-Cambrian Peneplain and areas with high relative relief and slopes within the exhumed Sub-Mesozoic relief.

In the south and southwest the Precambrian basement rises from below a cover of Mesozoic cover rocks. Here the topography of this sub-Mesozoic surface is characterized by high relief and consists mostly of undulating hilly land and structurally controlled landforms, especially along the west coast of Sweden /cf. Lidmar-Bergström 1996/. This implies that the Lower Palaeozoic cover, together with Devonian sediments, were eroded away during the Mesozoic. The relief of the sub-Mesozoic surface was developed by deep weathering during warm-humid phases of the Mesozoic, as evidenced by deep kaolinitic saprolites found at several places along the basement-cover rock contact in southern Sweden cf. /Lidmar-Bergström 1995, Lidmar-Bergström et al. 1997/.

In south central Sweden, on the other hand, the sub-Cambrian peneplain is replaced by a series of almost horizontal stepped erosion surfaces /Lidmar-Bergström 1988, 1996/. The major surface, the *South Småland Peneplain,* extends to 125–175 m.a.s.l. and truncates the exhumed sub-Upper Cretaceous relief to the south and is thus post-Cretaceous in age (Figure 6-2 and 6-3).

Quaternary ice sheets covered southern Sweden during the Elsterian, Saalian and parts of the Middle and Late Weichselian glaciations, but probably not during the Early Weichselian /Svendsen et al. 2004, SKB 2010, and references therein/. During the Late Weichselian the ice sheet briefly invaded the area from the north before it retreated northwestwards and the earlier ice sheets probably behaved in a similar way /Kleman et al.1997/. During most of the glaciations it is believed that the ice was frozen to the ground, but during certain glacial periods of time the whole area has probably been affected by erosive ice to some degree /Fastook and Holmlund 1994, Näslund et al. 2003/. In the northern and northeastern parts of the Laxemar study area, the occurrence of a thin or absent Quaternary cover /Fredén 1994; Figure 3/ points to warm-based erosive conditions during the deglaciation of the Late Weichselian ice sheet. /Björnsson 1937/ noted strong selective glacial erosion in the same area guided by the preglacial relief.

7 The Laxemar area

7.1 General outline

The Laxemar site is situated not far from the Baltic coastline on the eastern side of the South Swedish Dome. The bedrock in the area is dominated by c 1.80 Ga old intrusive rocks, which belongs to the Transscandinavian Igneous Belt and show variable composition ranging from granite to quartz monzodiorite to diorite-gabbro /Stephens and Wahlgren 2008/. The Quaternary cover is sparse in the coastal areas. Some 30 km inland from the coastline till becomes more abundant, i.e. above the highest postglacial coastline (Figure 7-1). Below the highest postglacial coastline, valleys are filled with silty clays and beach washed sands and gravel. Eskers are common and normally oriented NW-SE which shows the general direction of retreat of the Late Weichselian ice sheet over the area. From a geomorphological point of view the Laxemar area lies within a transition zone between the smooth till-dominated coastline south of Oskarshamn and the heavily dissected coastline to the north, see Figure 7-2.

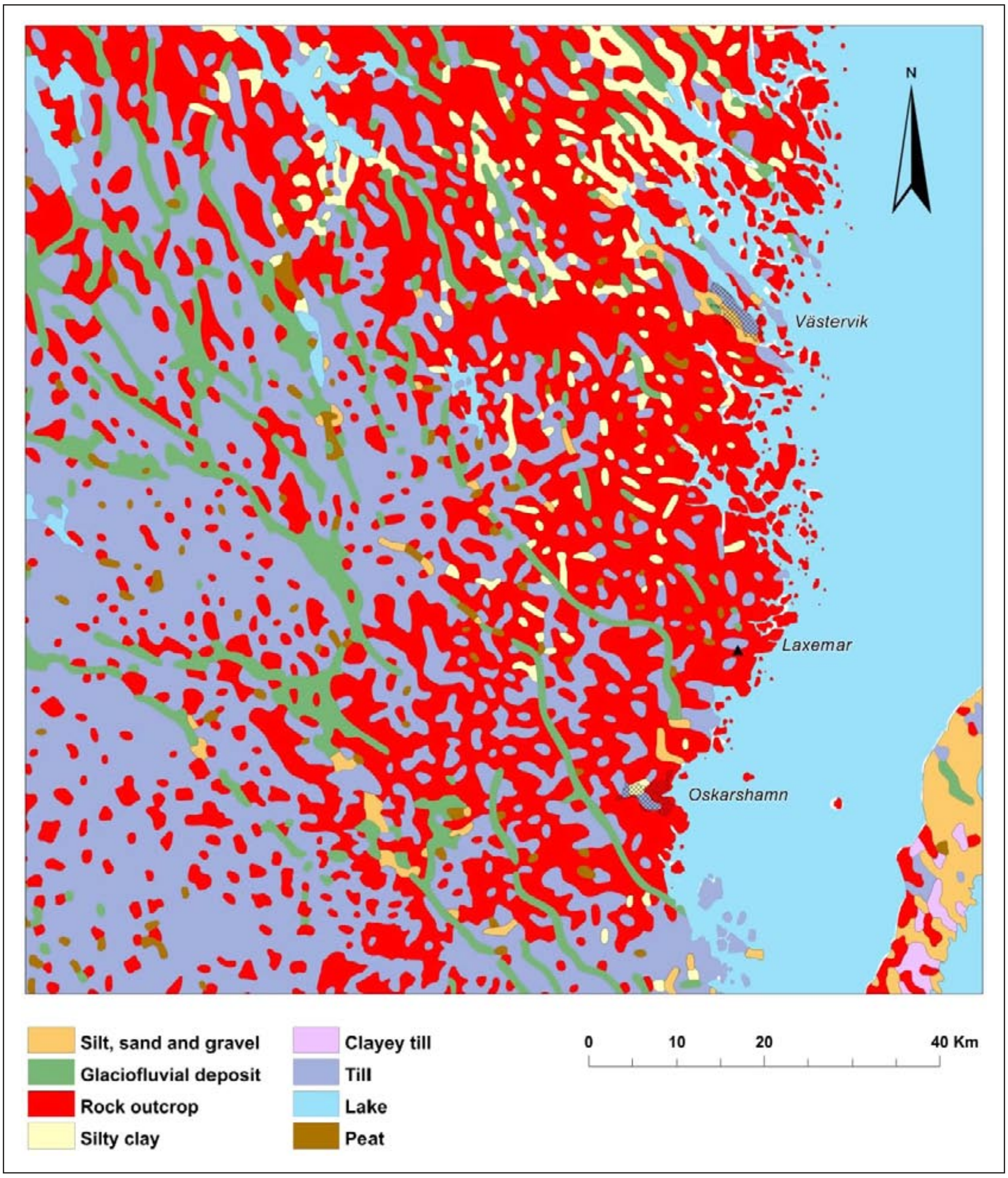

Figure 7-1. *Quaternary deposits in the Laxemar region.*

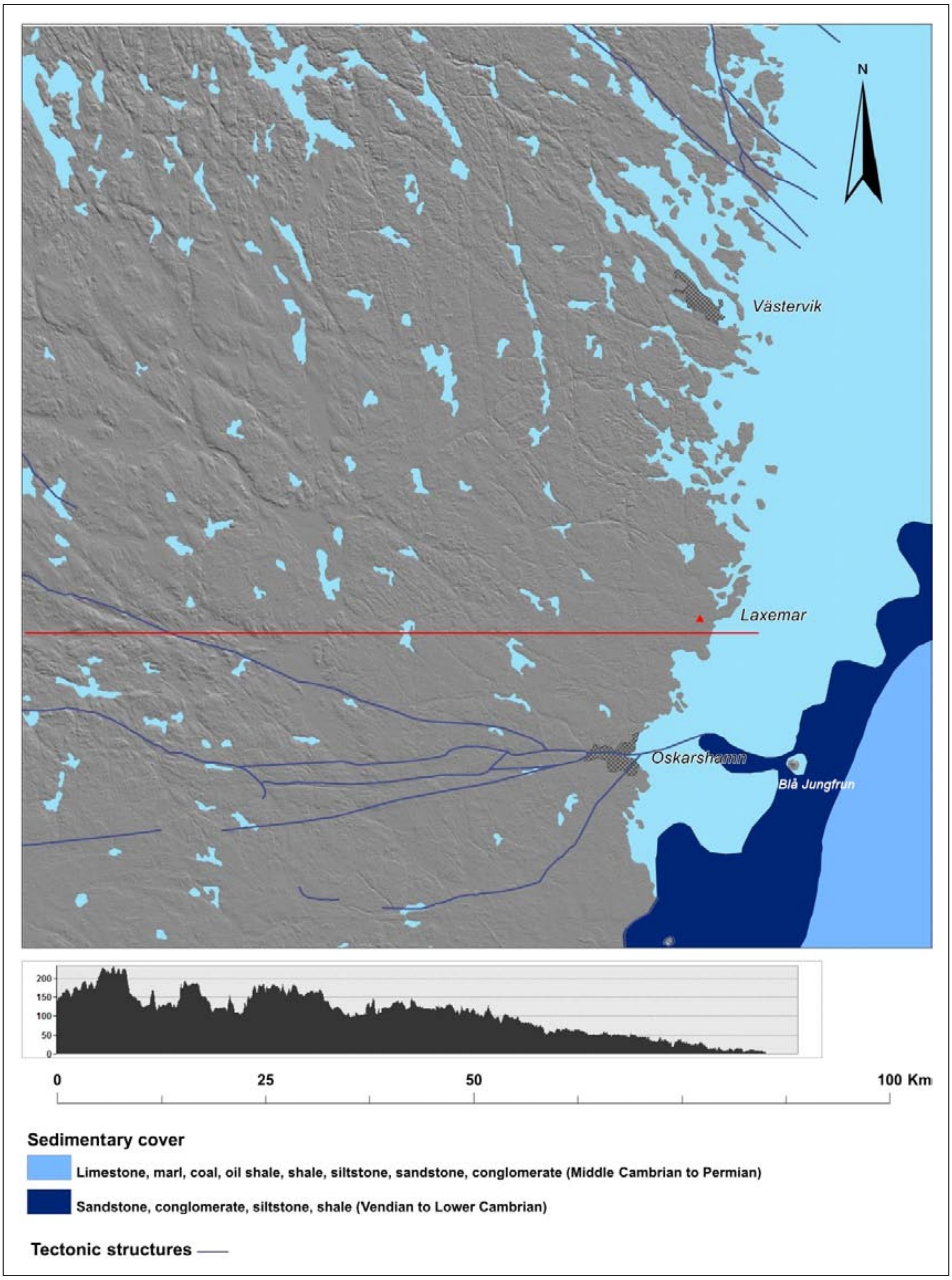

Figure 7-2. *Hillshaded relief map and profile in the Laxemar region showing the change in elevation and relief from the coast to the inland.*

In the entire area the land surface rises gently from sea level to close to 300 m.a.s.l. at a distance of 60–80 km inland. Here, the topography also becomes more dissected (Figure 7-3 and 7-4). The relative relief of the coastal areas in the southern part of the area is very low and normally within the 0–20 m range, while the relative relief in the coastal zone north of the Laxemar area often is between 20 and 50 m (Figure 7-5). Further inland and in the northern part of the mapped area the relative relief rises to above 50 m.

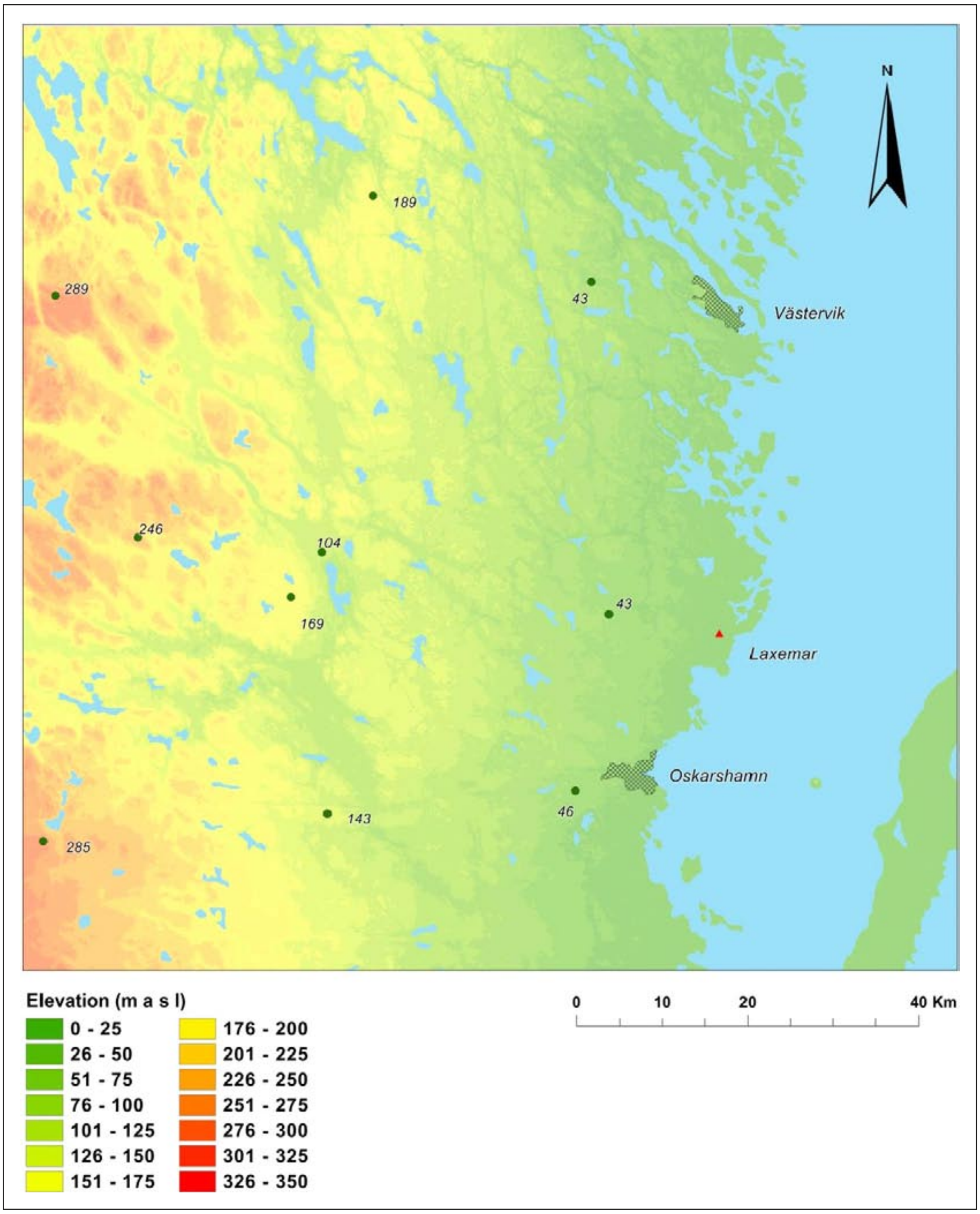

Figure 7-3. *Elevation map of the Laxemar area.*

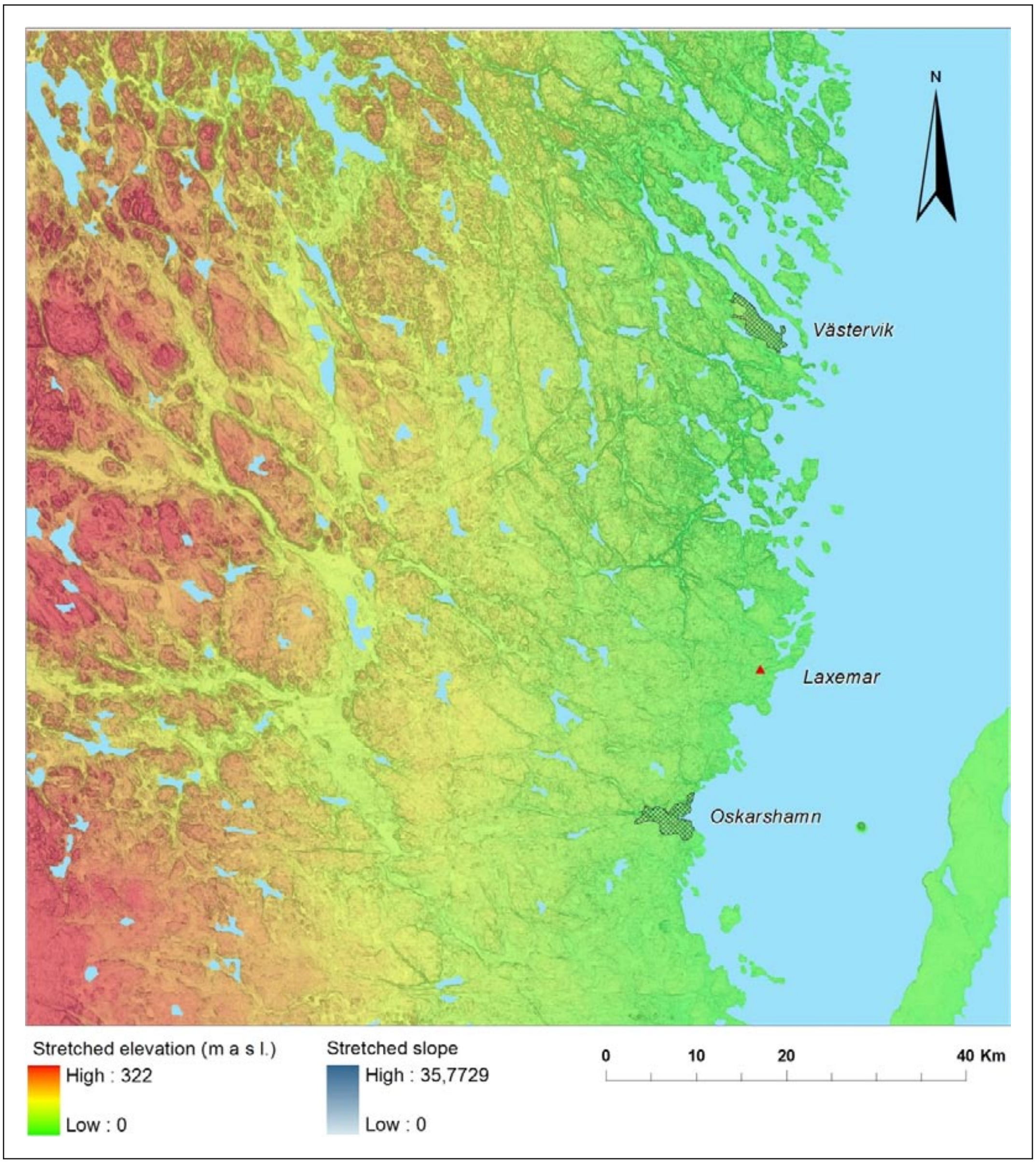

Figure 7-4. *Map of elevation and slope indicating the structural influence on the relief.*

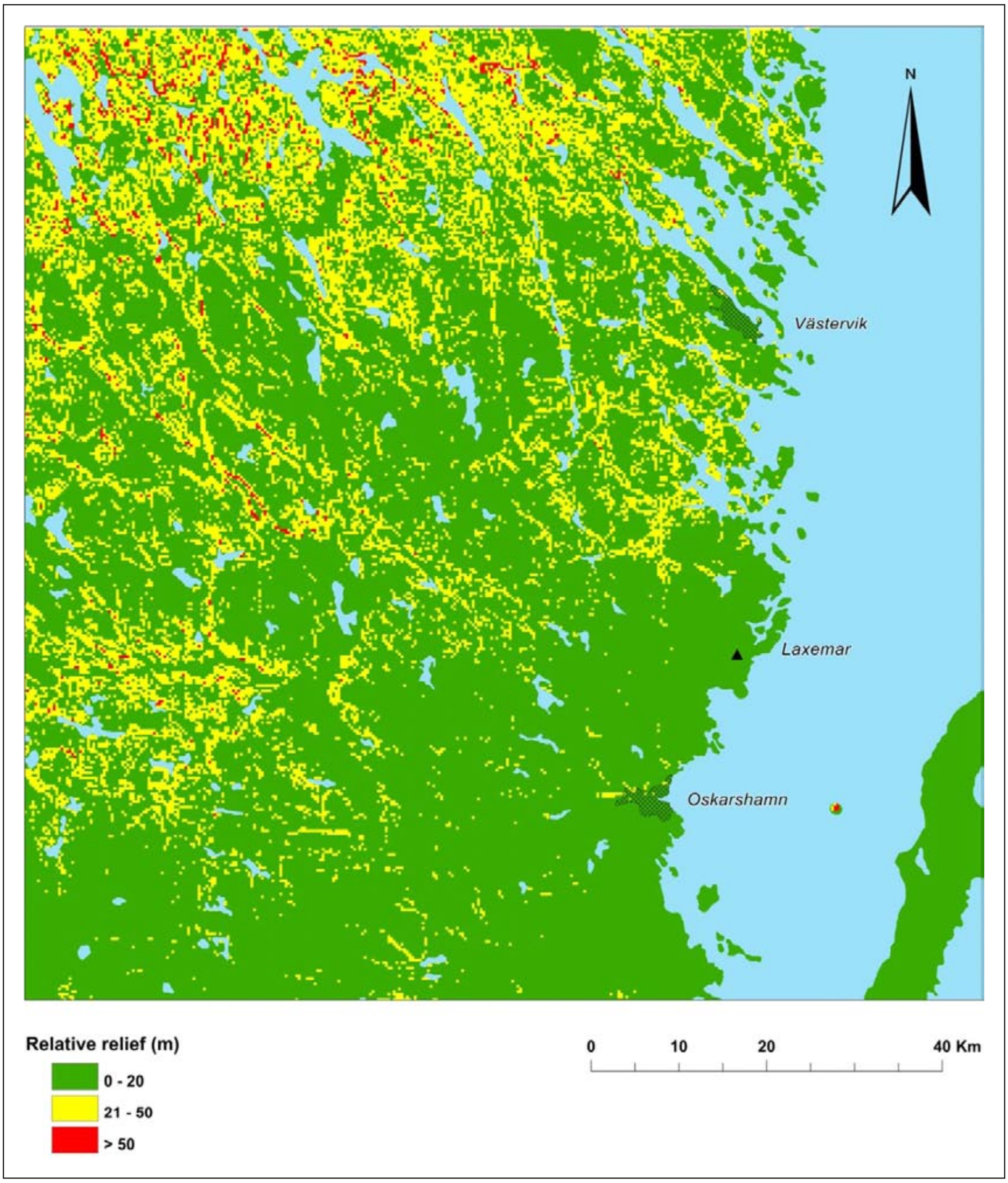

***Figure 7-5.** Map of relative relief. The very low relative relief along the coast south of Laxemar correspond to areas where the Sub-Cambrian Peneplain is intact.*

It should be noted from the map of relative relief (Figure 7-5) that a transition zone trending WSW-ENE occurs through the area. The area with higher relative relief meets the coastline just north of the Laxemar area and is coincident with a change in the outline of the coastline mentioned earlier. It is thus evident that the Laxemar area is situated within a well-preserved part of the sub-Cambrian Peneplain as defined by /Rudberg 1970/ and /Lidmar-Bergström 1995/. The relief in this part of the peneplain is astonishingly flat and it is notable that eskers appear quite clearly in the hill-shaded maps (Figure 7-6)

7.2 The peneplain at Laxemar – description and analysis

In detail the area surrounding the Laxemar site can be described as a landscape consisting of plateaux surrounded by narrow (50–200 m) shallow (10–25 m) valleys which follow the structural pattern in the bedrock trending WSW-ESE, WNW-ESE and SW-NE (Figures 7-6 and 7-7). The peninsula and embayments NE of Laxemar are outlined by these structures. The plateau surfaces are often very flat (Figure 7-8) and in some places apparently controlled by sheet structures (Figure 7-9). The importance of these structurally controlled patterns becomes more prominent 10–15 km north of the Laxemar site which marks the change to the more dissected surface in the northern part of the area.

***Figure 7-6.** Detailed hillshade map of the Laxemar region.*

***Figure 7-7.** Shallow valley typical of the Sub-Cambrian Peneplain in the Laxemar area. Photo from Bjurhidet. Photo Robert Hallström.*

***Figure 7-8.** Very flat bedrock surface characteristic of the bedrock plateau in the Laxemar area. Photo Robert Hallström.*

Figure 7-9. *Sheet structures control the shape of the plateau. Photo author.*

A key question is whether the plateaux and the valleys are integrated parts of the Sub-Cambrian peneplain or a result of denudation after the re-exposure of this old erosion surface. We know from other parts of southern Sweden that the relief of the peneplain in close contact to the cover rocks differs from area to area. These differences seem to depend on rock type. In southwestern Sweden very flat surfaces are developed in granites surrounding the dolerite capped mesas of Halle- and Hunneberg /Högbom 1910, Johansson et al. 1999/ while structurally controlled gneiss ridges characterize the peneplain close to the Palaeozoic outlier at Kinnekulle /Högbom and Ahlström 1924/.

The original shape of the peneplain in southeastern Sweden is more uncertain. However, an erosional residual, Blå Jungfrun partly covered by Lower Palaeozoic rocks (see Figure 7-2) is a prominent feature and evidence that the peneplain not was completely flat cf. /Lidmar-Bergström 1997/. It is therefore likely that the plateaux and structurally controlled valleys are situated at least very close to the original sub-Cambrian surface.

7.3 Glacial erosion in the Laxemar area

The total amount of glacial erosion in the area is very difficult to evaluate, however based on the conclusion that the sub-Cambrian peneplain is close to the present relief in this area it probably lies within less than 10 m of erosion in the bedrock. Glacial erosion is, however, responsible for the detailed shape of many bedrock surfaces in the area. This is especially notable in the coastal bare rock areas where glacial abrasion of stoss sides of rock bosses occur widely (Figure 7-10). Glacial excavation along pre-existing fracture zones is also indicated by the abrasion and plucking along the narrow embayments in the coastal areas (Figure 7-11).

Figure 7-10. *Glacial abrasion on stoss side of sheet structure at Bussviken. Photo author.*

Figure 7-11. *Features associated with glacial plucking and abrasion are common along the shores of the embayments in the Laxemar area. Photo from Långbonäs. Photo author.*

7.4 Evaluation of the Laxemar area

The Laxemar area is situated on the Sub-Cambrian Peneplian very close to the present extension of the Lower Palaeozoic cover. Exhumation of the peneplain has probably occurred during the Pleistocene glaciations and maybe as late as the Late Weichselian glacial phase. The major landforms in the area are interpreted as being very close to the original peneplain surface. Glacial erosion has reshaped some of the bedrock forms and may be responsible for evacuation of regolith along structurally controlled valleys. The total amount of glacial erosion in the area is probably less than 10 meters on average. However it is uncertain if this reflects the entire period of Pleistocene glaciations in the area. The very low relief of the landscape today together with the fact that the bedrock consists of crystalline basement rocks implies that the present denudation rate is extremely low and correspond to the low rates found in low relief shield areas reported in the literature, of the order of a few metres over a timescale of 1 Ma.

One key question is to what extent a new glacial cycle would change the landscape. We know that the peneplain further to the west was uplifted, weathered and dissected by fluvial and glacial erosion, probably starting in Late Pliocene or early Pleistocene /Lidmar-Bergström et al. 1997, Olvmo et al. 2005/. In these areas the relief is considerably higher if compared with the Laxemar area, of the order of 50–100 m. We also know that the Pleistocene glaciations in those areas have been incapable of removing all the weathering mantles that formed prior to the Pleistocene glaciations. Weathering during interglacial and interstadial periods is important since it controls how much weathered bedrock will exist before the onset of a new stadial/glacial. According to the section below (9), the denudation in the area during 1 million years is estimated to 0–5 m if there is no change in relief. This would imply that the effect weathering over timescales of 10–15 ka (expected duration of an interglacial) is quite insignificant. This is indicated by the fact that Southern Sweden has enjoyed interglacial for the last 9 thousand years or so and yet we still see striated surfaces.

8 The Forsmark area

The mapped area around Forsmark extends from the Mälardalen valley in the south to the city of Gävle in the north. As is evident from the map (Figure 8-1) most of the area is located below 100 m.a.s.l. The Forsmark site is situated in a coastal setting along the Baltic Coast approximately 60 km north of the city of Uppsala. Most of these coastal areas are below 25 m.a.s.l. A sharp boundary to higher terrain is visible in the west were the altitude quite rapidly rises to above 200 m.a.s.l.

The Quaternary deposits in the Forsmark region are dominated by till, which probably cover about 60% of the area (Figure 8-2). Clayey till occurs in some areas and is probably related to erosion of a former cover of Palaeozoic rocks that today are found in the Gävle embayment just north of the area. In addition, Riphean (Jotnian) sedimentary rocks occur in down faulted positions in the north in the Gävle Basin and in the east close to the coastline cf. /Lidmar-Bergström 1996/. This may indicate a very late exhumation of the Precambrian surface in this region. The distribution of bedrock exposures is interesting. An area with exposed bedrock can be followed from the Forsmark area to the Stockholm region in the south. This is probably partly a result of wave abrasion during isostatic uplift following deglaciation, but glacial erosion along the dissected coastline may also explain the distribution pattern.

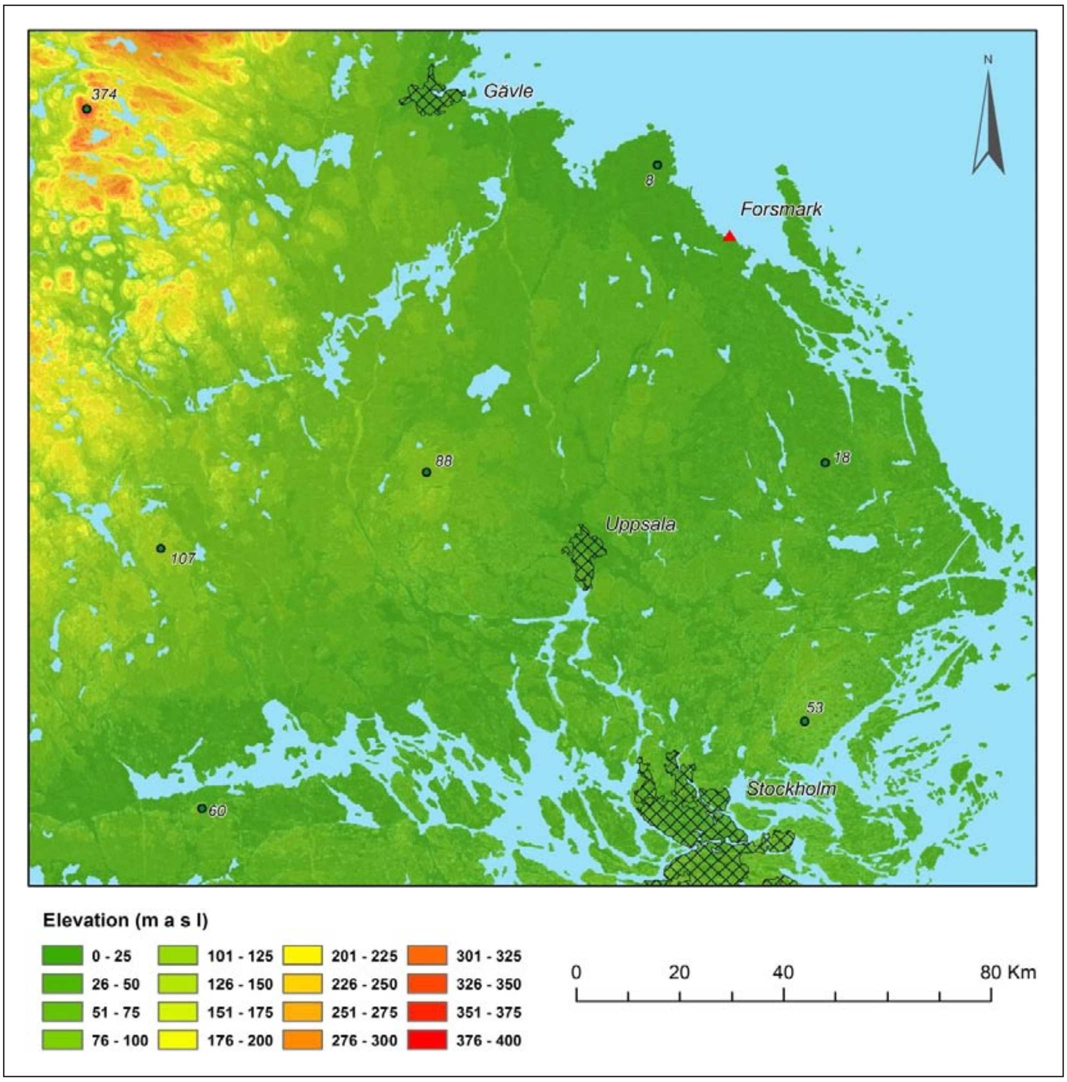

Figure 8-1. Altitude map of the Forsmark study area.

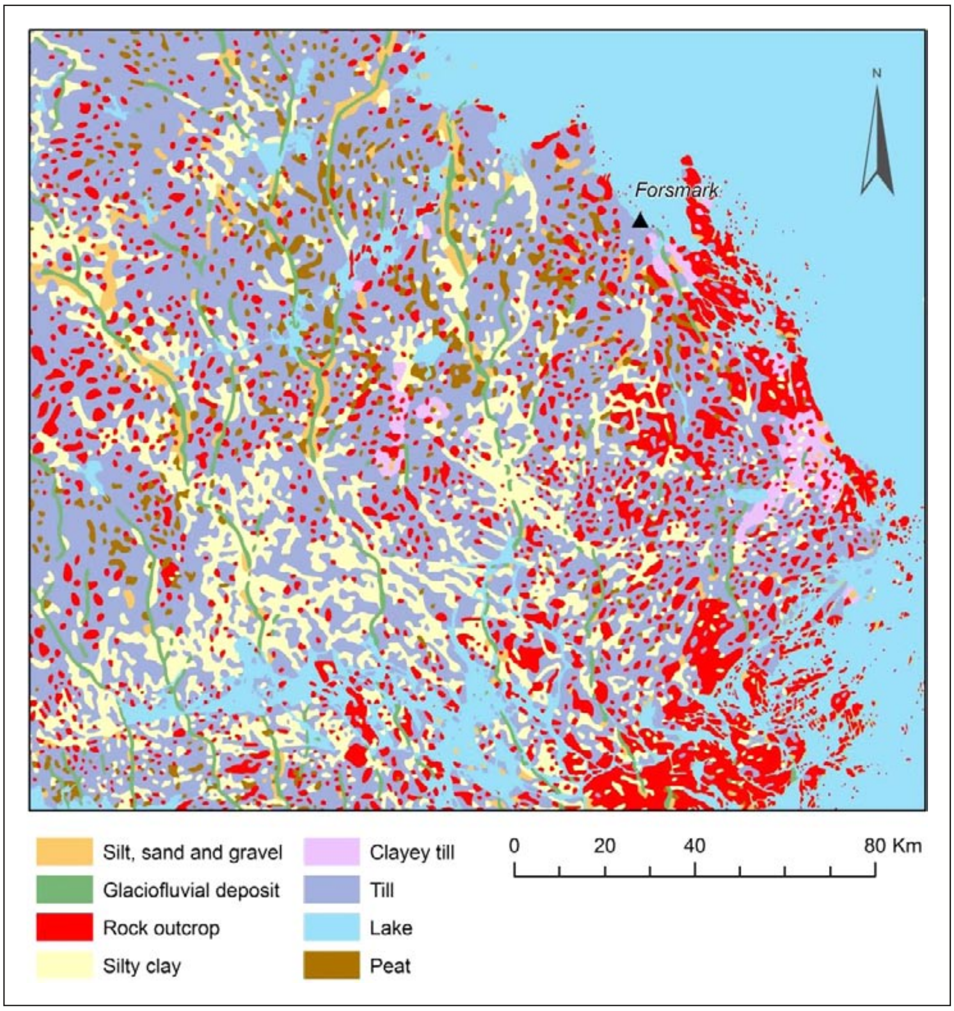

Figure 8-2. *Quaternary deposits in the Forsmark study area.*

Low lying areas such as basins and valleys are filled with fine grained Quaternary sediments consisting of silt, sand and gravel. Some very distinct and continuous esker systems are typical for this region and their north-south trend indicate the general direction of the ice recession during the Late Weichselian deglaciation. Some of these eskers are dominant features in the landscape, such as north of Uppsala.

From a geomorphological point of view the mapped area can be divided into four distinctive regions, 1) a central area characterized by very low relief (see Figure 8-3) , 2) a coastal region SE of Forsmark with coast parallel lineaments, 3) a southern part with more pronounced relief characterized by east-west trending fault scarps, and 4) hilly relief in the northwest. The Forsmark site is located in the first, central region of low relief.

The central part is extremely flat and in fact superimposed landforms in the Quaternary cover such as eskers are visible in the map showing the relative relief (Figure 8-4) as well as in the slope map (Figure 8-5). The relief in the central part, including the Forsmark site is characteristic of the sub-Cambrian peneplain and is consistent with the interpretation by /Lidmar-Bergström 1996/. The relative relief in this area is less than 20 m and the most remarkable geomorphological features are the eskers. However, some shallow valleys in the peneplain can be recognized in the slope map (Figure 8-5).

South of this central flat area, a more dissected landscape appears. Here the peneplain is broken into blocks and tilted in different directions cf. /Lidmar-Bergstöm 1996/. Some of the blocks are elevated and partly dissected by weathering and erosion. In the southern part of the study area, the elevated rims of these uplifted blocks give rise to E-W trending horst ridges, while in the Stockholm region the peneplain is highly dissected and give rise to a landscape consisting of plateaux bounded by shallow straight structurally controlled valleys.

***Figure 8-3.** The landscape between Uppsala and Forsmark are characterized by the sub-Cambrian peneplain with extremely low relief. Photo author.*

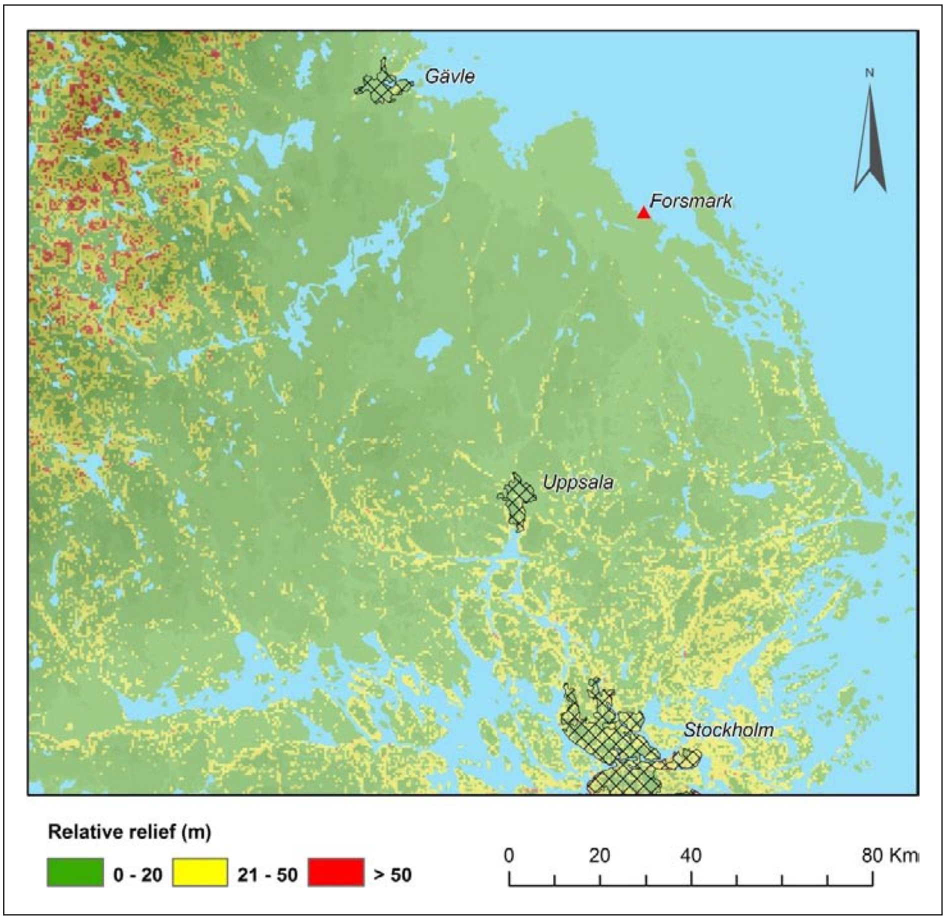

***Figure 8-4.** Relative relief in the Forsmark study area.*

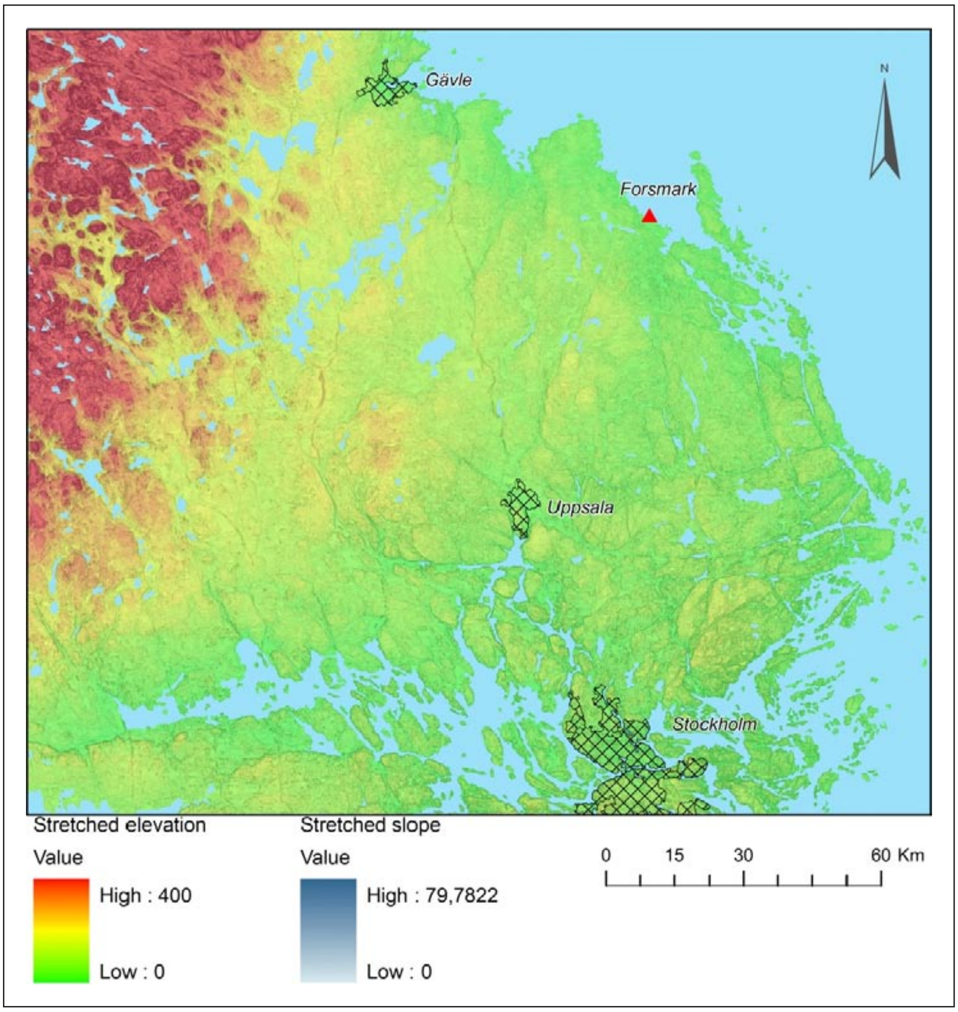

Figure 8-5. *Slope map of the Forsmark study area.*

The coastal low lying areas between Forsmark and Uppsala are of special interest. In this area the relief is quite considerable along coast parallel lineaments. This is obvious in the relative relief map (Figure 8-4) where relative relief locally rises to 50 m along these fracture zones. The slope map also points at a more dissected landscape in the coastal areas southeast of Forsmark. The area coincide with areas with a high frequency of rock outcrops which may suggest that glacial erosion have been more effective in this region, probably as a result of the closeness to the Baltic Sea depression which may have influenced the glacier flow. Glacially eroded rock surfaces area widely distributed along this coastline (Figure 8-6 and 8-7).

A more detailed map of the Forsmark area is seen in Figure 8-8, based on a hillshade map draped on an elevation map with stretched elevation values. The low lying flat coastal areas surrounding the Forsmark site is clearly visible both in the map and in the profile. The relative relief is extremely low and often below 10 metres. The very low relief of the entire area is further accentuated by the Uppsala esker, which is clearly visible both on the map and in the profile. A slight tilting of the peneplain towards the east is suggested in the profile and the slightly uplifted blocks appear as a couple of N-S trending elevated areas in the area south of Forsmark.

The structurally controlled pattern of islands, narrow straights and embayments are also clearly visible in Figure 8-8. Some of these morphological features coincide with tectonic lineaments outlined on the geological map (blue lines in Figure 8-8), but it is notable that not all of the morphological features forming linear features are marked as tectonic lineaments. The structural control on the geomorphology along the coastline is emphasised by the dissected coastline ca 25 km south east of Forsmark. This dissected area seems to coincide with the intersection of NW-SE lineaments and lineaments running approximately N-S. A large scale basin-like feature has been formed in the area of intersection. This is consistent with observations made by /Johansson et al. 2001/, which suggest that basins form at joint intersections.

Figure 8-6. *Deep structurally controlled coast parallel bedrock valleys with glacially eroded valley sides characterize the coastline southeast of Forsmark. Photo from Väddöviken. Photo author.*

Figure 8-7. *The low relief coast northwest of Forsmark is characterized by glacially eroded bedrock outcrops. Photo from south of Hammarsviksfjärden. Photo author.*

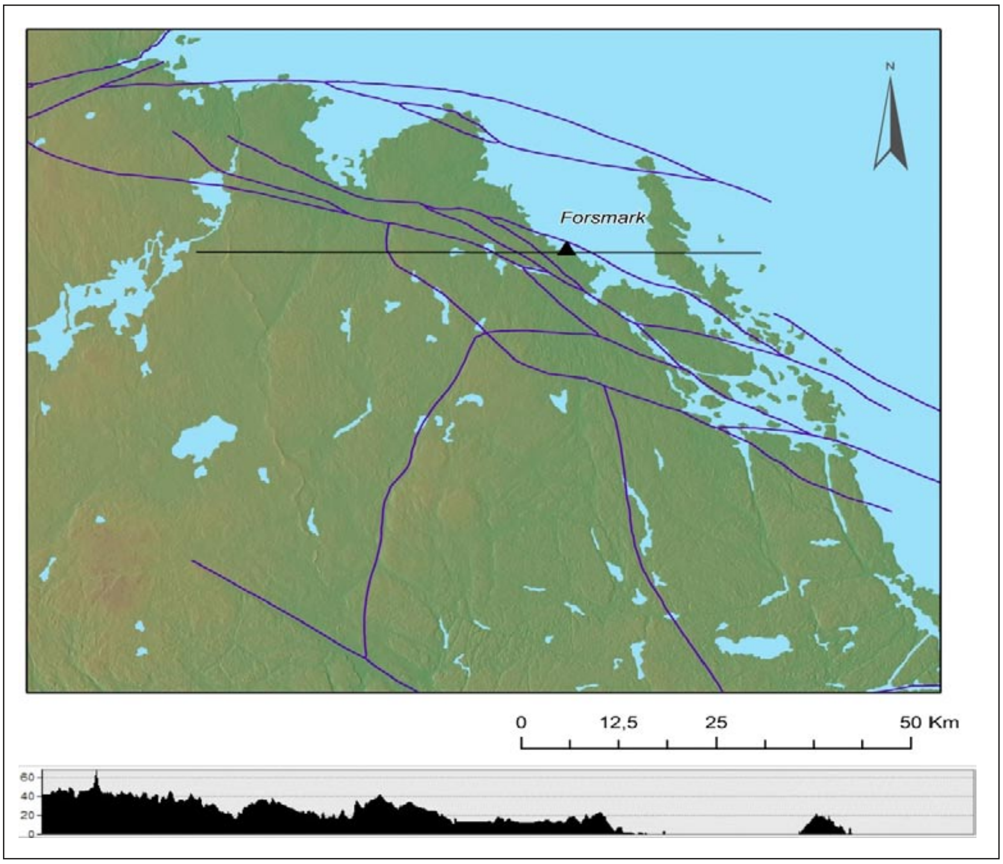

Figure 8-8. Hillshade map of detail of the Forsmark area. The hillshade is draped on a elevation map. Red colors are elevated areas. Green colors are low lying areas. Tectonic lineaments from the digital geological maps are included. Horizontal scale of profile is equivalent with map scale.

8.1 Evaluation of the Forsmark area

The Forsmark area is situated on the Sub-Cambrian Peneplain not far from the present extension of the Lower Palaeozoic cover. Exhumation of the peneplain has probably occurred during the Pleistocene glaciations and maybe as late as the Weichselian glaciation as indicated by the occurrence of clayey tills in the area. The major bedrock landforms in the area are probably close to the original peneplain surface. The area is extremely flat over large distances and many landforms in the Quaternary cover are prominent features in the landscape. Glacial erosion has reshaped some of the forms in the bedrock and may be responsible for evacuation of regolith along structurally controlled valleys. This is especially true in the coastal areas southeast of Forsmark, where glacial erosion may have been comparably effective along fracture zones. Based on the conclusion that the sub-Cambrian peneplain is well-preserved and probably coincide with the present relief the total amount of glacial erosion caused by repeated Pleistocene glaciations in the area on average is probably less than 10 meters, but is probably above that figure in the coastal zone southeast of Forsmark. The very low relief of the landscape today together with the fact that the bedrock consists of crystalline basement rocks implies that the present denudation rate is extremely low and correspond to the low rates found in low relief shield areas reported in the literature.

The effect of a new glacial cycle would probably be very limited in areas dominated by the well-preserved sub-Cambrian peneplain, such as at the Forsmark site, at least considerably less that the accumulated amount of glacial erosion during the Pleistocene. The main reason for this, compared to the cases of significantly higher glacial erosion rates presented in Figure 4-2, is that the bedrock relief and slope in the Forsmark region are very small, and that the expected magnitude of ice sheet erosion is smaller than the more effective erosion by mountain glaciers as presented in Figure 4-2. However, the dissected and sometimes considerable relief in the coastal areas may be a starting point for glacial erosion during a glaciation with a similar flow pattern as during the Late Weichselian. In this area glacial erosion of more than 10 m would be expected locally in low topographic positions.

9 Evaluation of the non-glacial contribution to denudation in a 1 million year time perspective

In order to evaluate the Forsmark and Laxemar regions in a 1 Ma time perspective, a potential denudation rate is calculated based on the relief in the two areas. The calculation is based on the functional relationship of /Ahnert 1970/ between denudation and relief. The calculation thus excludes the effect of land uplift and glacial erosion and simply relate denudation to the local relief assuming that the denudation rate D = 0.1535h, where *h* is the relative relief and is a substitute for mean slope. The maps of relative relief (Figure 7-5 and 8-4) in the two areas are used for the calculation.

For the Laxemar area (Figure 9-1) the calculation of potential non-glacial denudation give values between 0–5 m/Ma in the area around the Laxemar site. In the more dissected areas further west, the calculated values range from 5–10 m/Ma over large areas, while values between 10 and 15 m/Ma occur in the northern part very the relative relief is considerable. The calculated values are in good agreement with estimated denudation rates in this kind of geological and geomorphological setting from elsewhere.

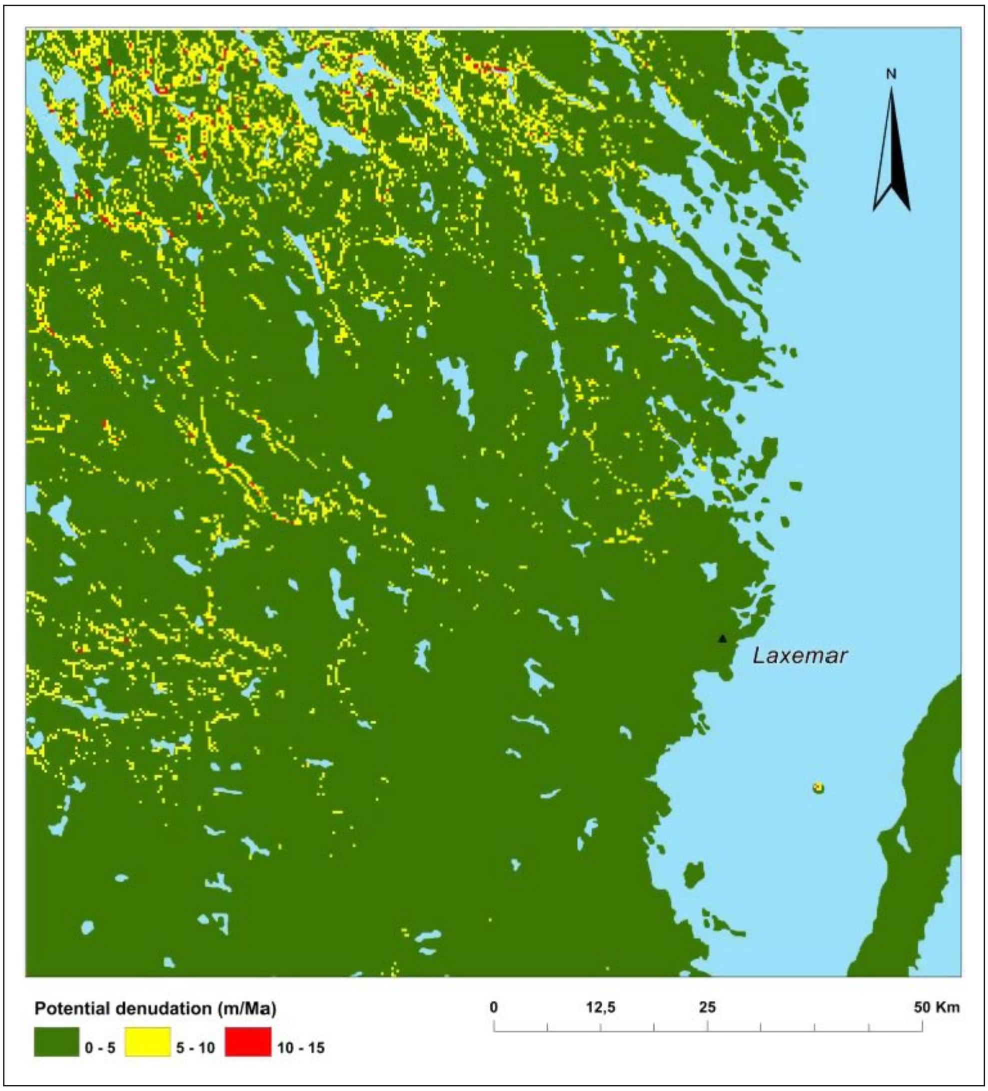

Figure 9-1. *Potential denudation in the Laxemar area.*

In order to show the effect of a change in relative relief by, for instance, tectonic uplift a map of potential denudation after a five-fold increase of the relative relief was produced (Figure 9-2). This means that the local relief in the area close to the Laxemar site rises to about 50 to 100 m. It should be noted that the calculation only involves a change in local relief, i.e. the local base levels, and not a changing sea level. In this case, the denudation rate in the area close to the Laxemar site increases to between 5 and 20 m/Ma. It also shows that the rate might increase to between 20 and 80 m/Ma in the dissected areas to the north. These figures are in quite good accordance with the denudation rates and relief formation induced by uplift of the South Swedish Dome in the late Tertiary cf. /Olvmo et al. 2005/.

The same sets of maps were produced for the Forsmark area (Figures 9-3 and 9-4). It shows that the denudation is between 0–5 m/Ma over most of the area, but might rise to between 5 and 10 m/Ma along uplifted fault scarps in the south and to 10–25 m/Ma in the northwest where the contemporary relief is higher. The case with a five time increase of relief makes no or little change in areas with an intact Sub-Cambrian peneplain, however, in the northeastern part of the study area the denudation locally rise to above 100 m/Ma (Figure 9-4).

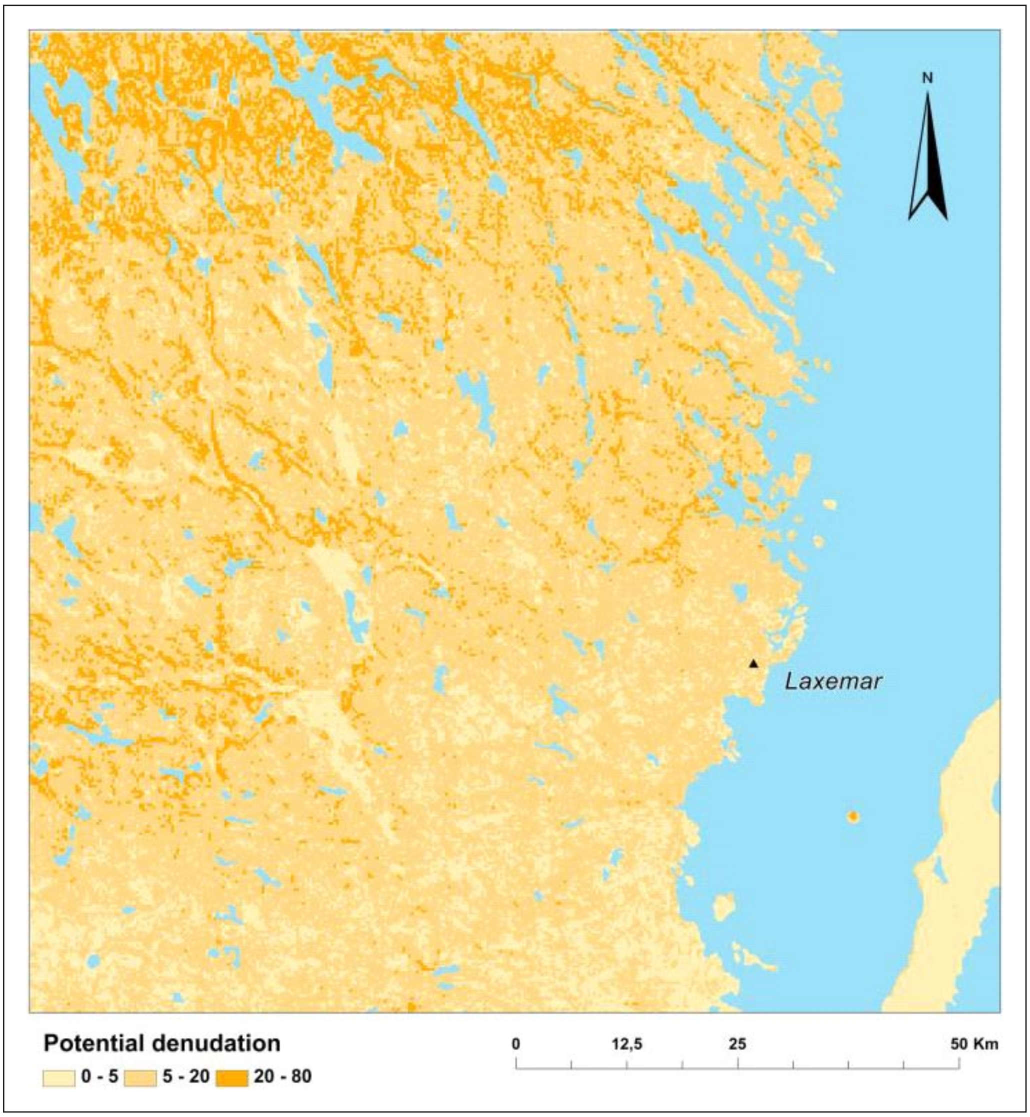

Figure 9-2. *Potential denudation in the Laxemar area with a fivefold increase in relative relief.*

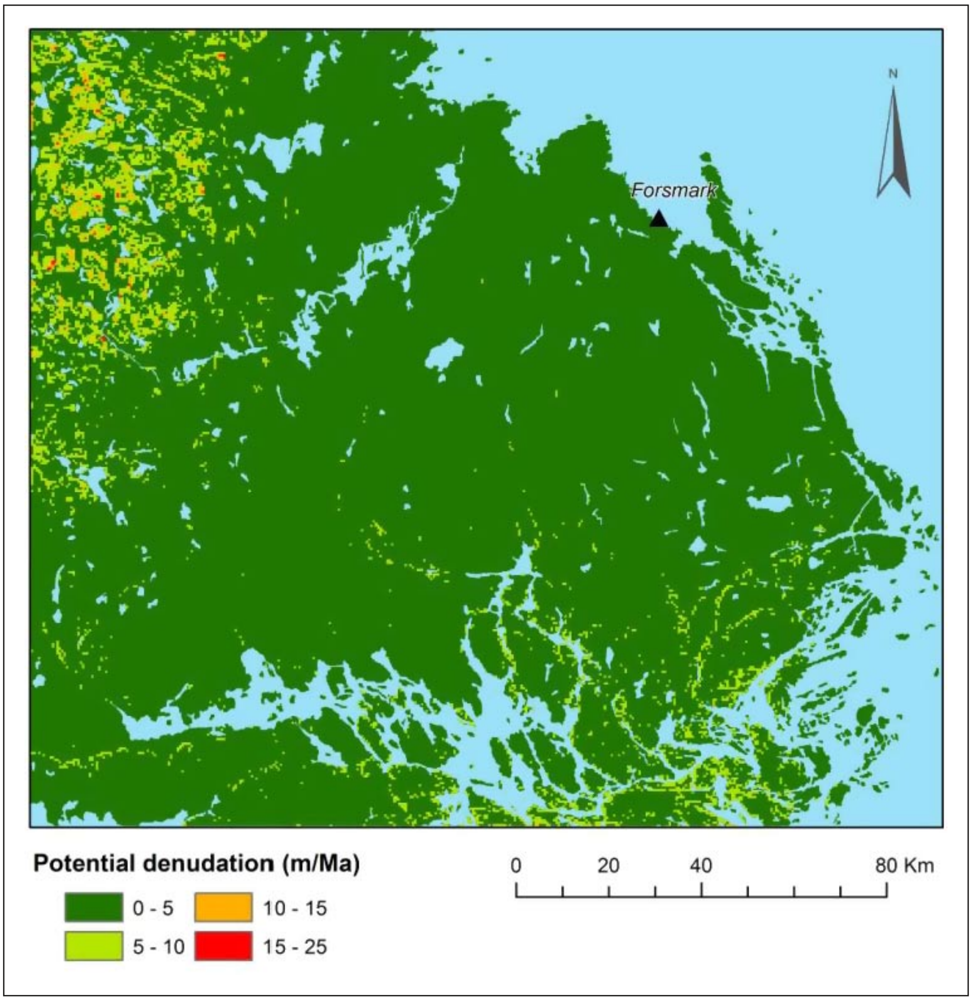

Figure 9-3. Potential denudation in the Forsmark area.

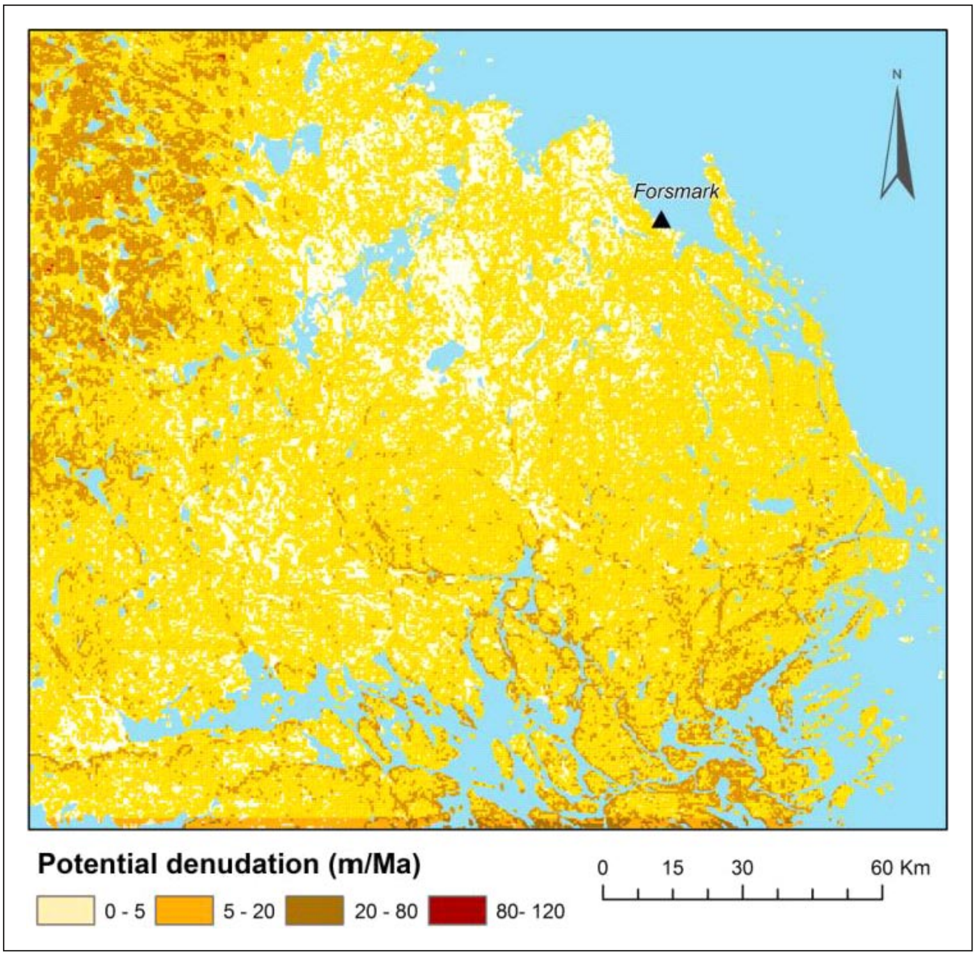

Figure 9-4. *Potential denudation in the Forsmark area with a fivefold increase in relative relief.*

10 Conclusions

The Forsmark and Laxemar areas are quite similar from a geomorphological point of view. Both sites are situated within intact parts of the Sub-Cambrian Peneplain with extremely low relief. The peneplain is considered to be intact at both sites, i.e. the present relief is quite close to the original surface of this old denudation surface. The total amount of glacial erosion up to present is estimated to be relatively low at both sites (less than 10 m on average), but may have been quite important for stripping of old regolith, especially along fracture zones. This is most obvious in the vicinity of the Forsmark site where the dissected coastline 15 km south of the site is interpreted as a result of glacial erosion along old fracture zones. Therefore, from a strict glacial erosion perspective the Laxemar site is somewhat better than Forsmark. However, given the presented long-term denudation up to present, the expected amount of glacial erosion during a future glacial cycle, if similar to the last glacial cycle, is probably very limited for both sites, in the order of 2–5 m.

Excluding glacial erosion, the long-term denudation rates of the two sites is fairly low as a consequence of the very low relief and the proximity to base level. The figures estimated for the long term denudation rates are in agreement with reports of denudation rates in shield areas and lie within the range 0 to 10 m/Ma for both sites. A scenario with a five-fold increase of relief that could be the effect of tectonic uplift, in a time perspective considerably longer than the coming 100,000 years, does not significantly change the picture. Even if the local relief is raised to above 100 m at both sites the estimated non-glacial denudation is very low. However, it should be noted that the effect of a relief change on the glacial system is not included in the calculations. Again, such a change on the pattern of glacial erosion would be more pronounced in the Forsmark area, since the coastline is this area seems to have been more exposed to glacial erosion than the Laxemar area.

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