Chapter 3
Glacial Erosion/Sedimentation of the Baltic Region and the Effect on the Postglacial Uplift

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Abstract Plio-Pleistocene erosion and sedimentation significantly impact postglacial uplift. We estimate in the last glacial cycle sedimentation could produce up to 155 m of subsidence and erosion 32 m of uplift. To show this we determine the changes in surface load caused by glacial and postglacial erosion and sedimentation over 1,000 year time intervals (coarser intervals before 50,000 years) utilizing a largely automated interpretation of regional geological and geomorphological observations that is constrained by plausible bounds on the rate of erosion of various lithologies and the known general pattern and behavior of glacial ice (ice boundaries over time, the dendritic pattern of ice movement, geometry of fast-flowing ice streams, plausible changes in frozen-bed conditions, etc.). Mass balance between erosion and deposition is enforced at all times. The analysis is regional and obliged to agree with all known geological constraints. Although the focus is on the last glacial cycle, all previous cycles are considered. The analysis suggests that the first glaciations probably shaped the major overdeepened troughs, although it is possible that the deepening was distributed evenly over all the cycles. Younger glaciations mainly removed sediments left by their predecessors, decreasing the thickness of the Quaternary succession and only locally incising and changing the dip of the bedrock surface. Over the last glacial cycle, ~20–90 m of sediments (and locally more) was removed in the zones of most active erosion.

Keywords Pleistocene · Glaciation · Erosion · Sedimentation · Isostasy · Fennoscandia · Baltic · Ice-stream · Uplift · Bedrock

3.1 Introduction

The role of glacial erosion and sedimention in creating the modern landscape of the Baltic Sea basin has been appreciated for a long time. Glacial and fluvioglacial erosion had a decisive influence in shaping the Baltic–White Sea lowland on the

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margin of the Fennoscandian (Baltic) shield (Amantov 1992, 1995), for example. The Atlantic margin shows increased Late Pliocene and Pleistocene deposition rates (Riis and Fjeldskaar 1992). Worldwide, erosion of exposed unconsolidated clastic shelf sediments and consequent isostatic compensation has resulted in large masses of sediment being offloaded from the continental shelves onto deep-sea fans and abyssal plains by turbidity currents (Hay 1994). But opinions differ on the intensity of the glacial erosion. To some the glaciations were crucial in changing the landscape. These authors emphasize that glacial erosion can be much greater than fluvial erosion (White 1972, 1988, Bell and Laine 1982, 1985, Clague 1986, Braun 1989, Harbor and Warburton 1992, 1993, Clayton 1996, Hallet et al. 1996, Montgomery 2002, James 2003). In mountain glaciers, the erosion rate is greatest near the equilibrium line altitude (ELA) where ice accumulation changes to melting. Here the glaciers are often considered "buzz saws" (Brozovic et al. 1997, Meigs and Sauber 2000, Montgomery et al. 2001, Mitchell and Montgomery 2006). Glaciers increase topographic relief through a combination of focused erosion in valleys and the regional isostatic rebound the incision induces (Small and Anderson 1998), and this, in turn, increases erosion.

Other researches point to the moderate transformation of preglacial landscapes and find evidence for low rates of glacial erosion and little difference between fluvial and glacial erosion rates (e.g., Gravenor 1975, Sugden 1976, 1978, Lindström 1988, Hebdon et al. 1997). In this view, the glaciers merely polished the northern shields, and the erosion they caused (although sometimes highly variable; Lidmar-Bergström 1997) was generally less than tens of meters in magnitude.

Glacial erosion is intriguing because on the local scale it is highly irregular but at the large scale it is regular. We would like to understand it quantitatively. For example, we would like to assess whether most of the sediment redistribution took place during the first or last glacial cycles. The shifts of sediment loading could be enough to affect subsurface temperature and cause isostatic tilting. But local spatial variations, the wealth of data that must be assembled and integrated, and the large spatial scales involved make analysis difficult.

Our approach is to apply computer software adept at creating and manipulating surfaces to infer glacial erosion and sedimentation rates across Europe in a locally detailed but regionally coherent way. At every instant of time and across the Quaternary, our method requires that erosion and sedimentation are balanced, locally and across all of Europe. Our analysis honors bounds on what erosion and sedimentation rates are reasonable, and a great many local geological constraints. The redistribution is process-driven. We develop algorithms that honor the pattern of glacial flow suggested by geological evidence and the locations of the ice margins as the glaciers grew and retreated. From this we build erosion and sedimentation modules that redistribute the sediments. We calibrate these tools to current glaciers and to the observed present-day sediment pattern, and this assures they are reasonable (but not necessarily correct). In this manner, we infer how the sediments may have been created and redistributed across Quaternary time and tentatively conclude that most of the major bedrock landscape changes were probably produced by the earliest glaciations. Even so, the erosion and sedimentation that occurred over the last glaciation were still sufficient to induce isostatic movements comparable to those caused by glacial loading. The analysis suggests interesting phenomenological connections.

The purpose of this chapter is to present and describe the results of this new kind of analysis. We motivate the methods used with a fairly extensive review of geological observations that provide insight into the processes that are occurring and the parameters that appear to be important and then give a fairly brief discussion of the methods themselves. The results of our analysis are then given in reasonable detail.

Although physically based, our methods remain largely empirical algorithms. As such they are difficult to fully describe in any reasonable space, and, in any case, their validity rests largely in their predictions. We will describe the methods in full detail in subsequent publications. Here we hope mainly to show that the sediment redistributions that result from the analysis we describe are reasonable and interesting.

3.2 Glacial Erosion and Sedimentation

Rates of glacial erosion have been estimated between 0.1 and 10 mm/year. Erosion of glaciated catchments of fjords of southern Alaska exceeds 10 mm/year (Hallet et al. 1996). Long-term averaged exhumation rates are 3 mm/year in the Chugach–St. Elias Range, Alaska, where the maximum rates of denudation are thought to be limited by rates of tectonic uplift (Spotila et al. 2004). In Western Nunavut, 6–20 m of rock is believed to have been eroded during the last glacial cycle (Kaszycki and Shilts 1980).

In Northeast Scotland, where both glacial and preglacial landforms exist in close proximity, the expansion of ice sheets across the area in the middle Quaternary was associated with a sharp increase in the rates of erosion (>0.13 mm/year), but the last (Late Devensian) ice sheet in the area was less erosive (<0.095 mm/year) (Glasser and Hall 1997).

On the assumption that the erosional work was achieved over 10,000–20,000 years of each 100,000 year glacial cycle, the rates of surface lowering during glaciations in Britain fall in the range of 11–23 mm/year (Clayton 1996). The average erosion rate over the full glacial cycle is comparable to the 1 mm/year figure regarded as "typical" by Boulton et al. (1991) for glacial erosion of resistant rocks. Erosion in Britain is several times faster for weaker rocks flooring major lowlands and much of the shelf (Clayton 1996). Average erosional rates of the sedimentary bedrock of the Barents Sea during the last ~2.5 million years were estimated to be between 0.1 and 1.1 mm/year (Faleide et al. 1996) or 0.2–0.6 mm/year (Solheim et al. 1996).

Assuming glacial erosion for 1 million years over the past 2.57 million years, the average rate of glacial erosion in the Sognefjord drainage basin, western Norway, was ~0.4 mm/year by subtracting the present topography from a reconstructed preglacial (paleic) surface. Considering the selective nature of glacial erosion along ice streams, the annual erosion rate for ice streams is most likely 2 ± 0.5 mm/year (Nesje et al. 1992, Nesje and Sulebak 1994). Comparable mean rates were reported for Isfjorden region of Svalbard over the glacial cycle (Elverhøi et al. 1995). In the Antarctic, average erosion rates are considered to be three times higher beneath ice-stream tributaries which are underlain by deep subglacial troughs (0.6 mm/year) than beneath ice-stream trunks (0.2 mm/year) (Bougamont and Tulaczyk 2003).

Remnants of marine Quaternary sedimentary sequences indicating high glacial/fluvioglacial erosion rates in the Baltic–White Sea lowland are a cornerstone in the validation of our erosion–accumulation modeling. The sporadic distribution of the youngest marine interglacial strata in the form of remnants around the Gulf of Finland attests to strong erosion during even the last glaciations (which was the smallest of all in this area). Sediments from previous glacial cycles are very rare in the axial part of the lowland, but in rare isolated locations remnants can be 40 m thick (Malakhovsky and Amantov 1991). Surface reconstruction suggests that, in addition to thick marine strata, at least 10–20 m of the underlying sediments were removed. In ice-stream zones like Lake Ladoga, remnants of older Quaternary beds survived the deep erosion in protected positions, indicating more than 60–70 m of erosion during the last glaciation. This suggests that in zones of active erosion the present cover belongs nearly entirely to the last glaciations (moraine cover and late-postglacial sediments).

Where soft sedimentary sequences have been glaciated, buried channels and hollows of several generations suggest local linear erosion of 100–200 m (Amantov 1992). Rarely, older channels can be seen to be entrenched at shallower depths than the younger channels that cross them (Amantov 1992). The nature of these channels depends on whether they are radial or parallel to the glacial front, affected by sedimentary infilling, deformed by ice or melting waters, etc. Lithology and structure are also dominant factors. The channels may often have nearly parallel orientation, sometimes with arc shape that roughly coincides with the boundaries of retreated glacial tongues. The depth of the channels decreases in the direction to the modern shield, so that the base of the channels tends to parallel the relief of the basal platform sediments, mostly entrenching only into the weathered top of the resistant crystalline basement. A similar rapid decrease in channel depth occurs toward resistant lithologies such as carbonate rocks forming prominent scarp-like features on the bedrock topography.

The depth of both glacial and fluvioglacial erosions strongly depends on lithology (Amantov 1992, 1995). In the Baltic–White Sea, depressions in bedrock topography suggesting maximum long-term erosion are evident in zones with pliable sediments. Here, glacial erosion rates inferred geologically and in our analysis reached 2 mm/year, with local short-term rates up to 8 mm/year. The thickness of erodable sediments should be taken into account. The rate of erosion should decrease if a pliable sedimentary unit is completely removed in an area with exhumation of resistant surface. The Landsort Deep illustrates how removal of a thickness of pliable sediment can create a strongly overdeepened ice-proximal negative form.

Another key factor controlling glacial erosion is the ice sliding velocity at the ice-bed contact (Humphrey and Raymond 1994). Our analysis addresses ice-stream zonation and accounts for the radial increase in ice velocity outward from the central zone of ice accumulation to the abrasion maximum near the ice terminus. In our models, abrasion increases up to a point and then possibly decreases due to overwhelming of the abrasive content that reduces basal sliding velocity by increased basal friction. Ice boundaries thus control concentric changes of the erosion rates. This broad pattern provides a regional context for further refinements. The main refinement in the erosion pattern is caused by fast-flowing ice streams near the glacial margins that have an enhanced capacity for erosion (Fig. 3.2). Ice streams move at high velocities under low driving stresses in a basal zone environment mostly because their base is lubricated (see discussion in Marshall et al. 1996, Tulaczyk et al. 2000, Stokes and Clark 2001, Kamb 2001, Bougamont and Tulaczyk, 2003, Hall and Glasser 2003).

The bedrock surface determines the topography of ice streams with profound erosion capacity. The location of bedrock troughs or elongated lowlands was initially controlled or at least influenced by the bedrock topography. Domination of elongated landforms of smaller scale is taken to indicate zones of faster ice flow. The elongation ratio of bedrock forms and megascale lineations are known to be a useful proxy for ice velocity (Anderson and Shipp 2001). Long subglacial bedforms (length:width ratios 10:1) are indicative of fast ice flows (Stokes and Clark 2002). The geological–geomorphological impact of ice streams cannot be underestimated, since modern ones literally control ice discharge. For example, over 90% of ice discharging from the West Antarctic Ice Sheet into the Ross Ice Shelf (Joughin and Tulaczyk 2002) is carried by ice streams.

Bedrock surface forms may also suggest very low ice velocities and erosion. Areas with abundant distribution of relict landforms indicate slow ice. Special grid filtering to emphasize outliers with a relevant search window can identify these areas best. In zones adjacent to weathered bedrock, possible frozen-bed conditions and weak erosional capacity can be manually input as constraints.

3.3 Methods

The preceding section suggests what must be taken into account by any glacial erosion analysis. Not discussed thus far is that the mass of glacial sediments must equal the mass of material eroded. We compile a huge quantity of published seismic and sedimentological data and make our best estimates of the total sediments deposited across the Quaternary. This provides a bound on the total Quaternary erosion. We use denudation surfaces to estimate the erosion directly. This stage of analysis is essentially an automation of traditional methods (Riis and Jensen 1992). Surfaces capture stages of Tertiary uplift and erosion (Amantov 2007). The surfaces connect isolated summit outcrops, patches of exhumed peneplains, and etchplains. Surfaces emerging from under sedimentary cover can be extrapolated and correlated with onshore saprolites and (or) remnants of cover so that the grids measure missing volumes. The surfaces can also illustrate past geological conditions. Regional compilations always have some uncertainties due to gaps in confirmation of seismic stratigraphy, different estimations of drainage provinces, and possible input of eroded material from irrelevant provinces to depocenters, etc. We estimate that the amount of material eroded in the Baltic region during Plio-Pleistocene is about 90,000 km^3 (Amantov 1995).

We estimate both the erosion and the sedimentation over specific intervals of time and require that erosion equal sediment accumulation over these periods. We use 1 ka timesteps over the last 50,000 years and longer 5–10 ka steps for early Weichselian stadials and across earlier glacial cycles. For the early Weichselian we assume two interstadials with ice-free conditions following Lundqvist (1992) and Lokrantz and Sohlenius (2006) as corrected by Svendsen et al. (2004) and Sarala's (2005) interpretation for southern Finnish Lapland.

The margins of the glacial ice sheets are the starting point for our analysis. The ice margins at the LGM are shown in Fig. 3.1. We use a number of tools to simulate erosion under the ice cover and sedimentation under, at the margin, and outside the ice. The tools are computation modules that allow useful geological analysis procedures to be repeated easily. The procedures might include sampling of gridded data (sub-ice lithology, for example), connecting sparse kinds of data with a best fitting surface, inferring velocity fields from the distance to an ice depocenter and topography, subtracting surfaces to determine the material removed, visualizing the geology in particular ways, etc.

Erosion under the ice sheets is estimated using such tools by requiring that the long-term glacial erosion rates are reasonable and the pattern of erosion conforms to the concentric (radial) changes in erosion observed as well as the "spider's web" pattern of grounded ice sheet's movement (ice streams). This is illustrated in Fig. 3.2. Figure 3.2a shows the erosion and sedimentation that might occur if only the ice velocity were considered. The concentric pattern results from the low ice velocity under the center of the continental glaciers and the more rapid basal ice velocity near the margins. Figure 3.2b shows how this simple pattern is modified if the likely effect of the spider-web pattern of ice flow with the enhanced erosional capacity of ice streams is taken into account. Figure 3.2c illustrates the effect of different erodability of sedimentary bedrock and basement lithologies. The glacial erosion module contains adjustable parameters that allow the sediment redistribution it "predicts" to be controlled by only concentric factors (Fig. 3.2a) or increasingly influenced by lithology, dendritic ice flow, and ice streams (Fig. 3.2b, c).

An important control is sub-ice topography which helps control the spider-web flow with "topographic" ice streams and erosion paths. The drainage pattern is determined from the paths raindrops runoff would follow in reaching the sea. Submodules refine the interpretation. For example, overdeepening of bedrock surface is imposed where slopes are >10–20° and oriented such as to cause rotational ice flow that could locally maximize basal sliding (Lewis 1949). The modules create grids that capture erosion surfaces over time and show the exhumation of sedimentary rocks, the boundaries of the sedimentary cover, expansion of the crystalline shield exposure, etc.

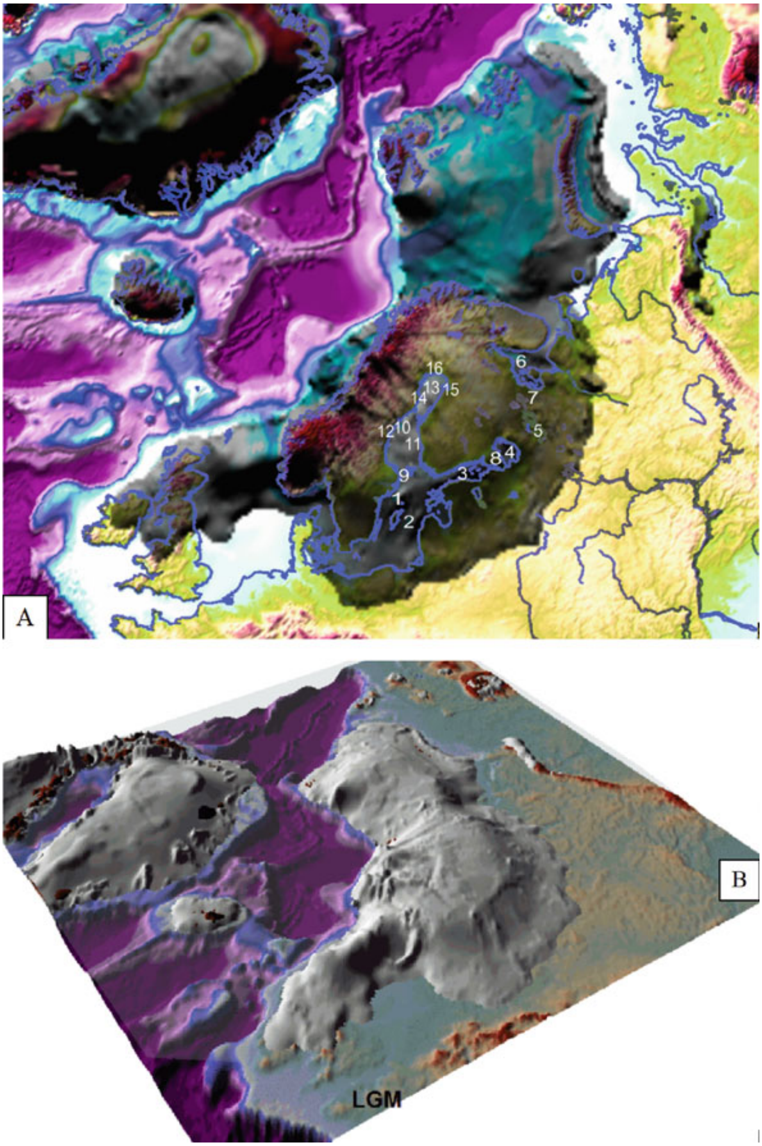
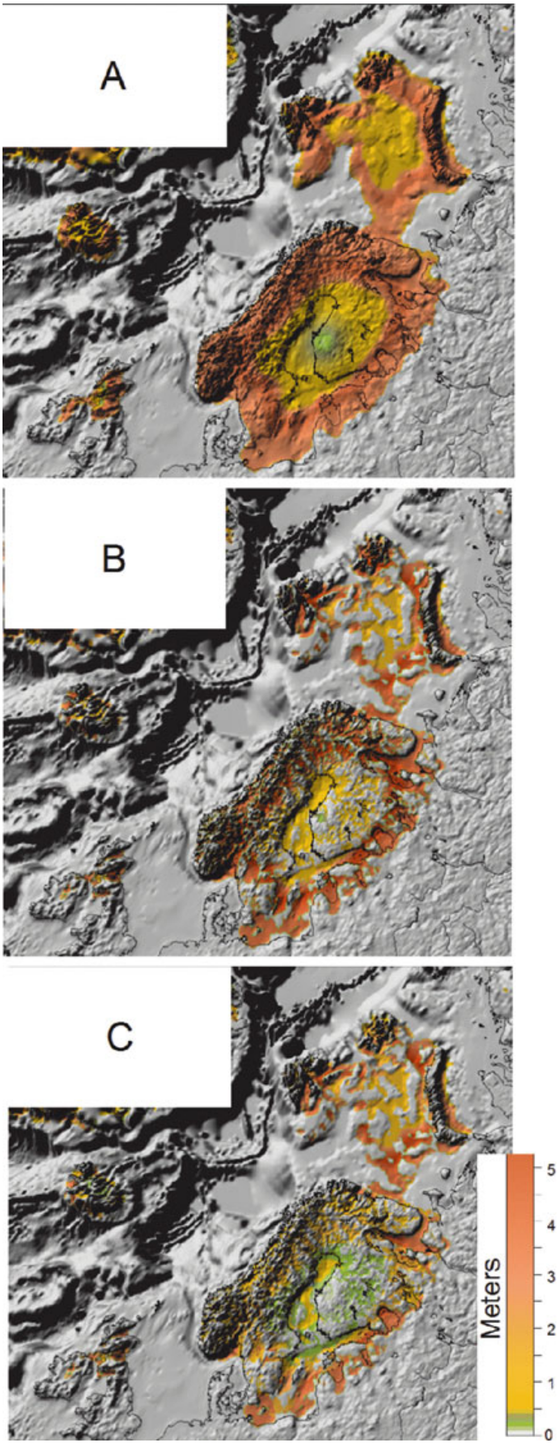


Fig. 3.1 Sample output of ice thickness module (LGM): **a** Orthographic view, ice is shown in half-transparent mode. Present-day shorelines are shown in *blue color*. Figures illustrate localities mentioned in the text. Central Baltic Proper: *1* – Landsort Deep, *2* – Gotland Deep; *3* – Gulf of Finland; *4* – Lake Ladoga; *5* – Lake Onega; *6* – White Sea; *7* – Vetryany Poyas; *8* – Karelic peninsula; *9* – Åland Deep; Bothnian Sea: *10* – Hörnösand Deep; *11* – Aranda Rift; *12* – Sundsvall; *13* – Bothnian Bay; *14* – Shellefteo; *15* – Ouly; *16* – Nordkalott. **b** 3D view

Fig. 3.2 Sample output of glacial erosion module: routine transformation from general simplified concentric pattern (**a**) to ice-stream flow (**b**) and further account of different lithology and erosion resistance (**c**)



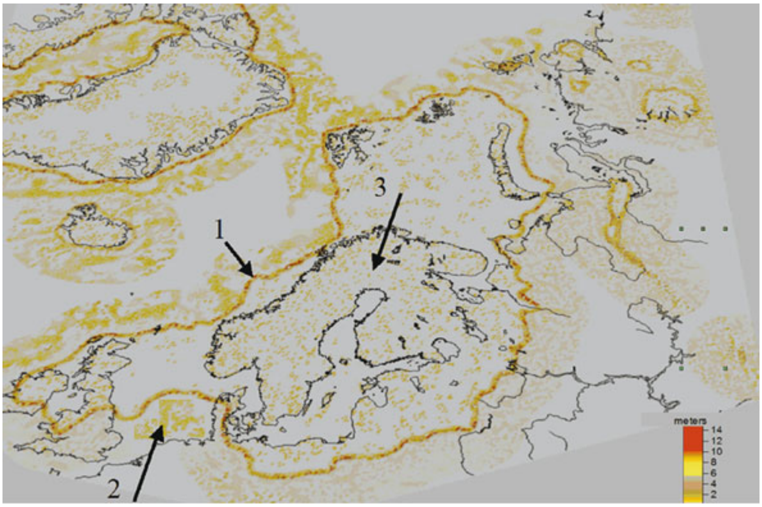


Fig. 3.3 Sample output of glacial sedimentation module: *1* – end-moraine ridges; *2* – peripheral sediments; *3* – products of subglacial sedimentation

The complement to erosion is sediment accumulation. This can occur under or outside the ice. We distinguish the glacial, interglacial, and postglacial sediment deposition patterns. For example, a *glacial sedimentation module* simulates the formation of end-moraine ridges, subglacial (i.e., drumlins, flutes, eskers), and peripheral deposition that deals with meltwater redeposition of a material (Fig. 3.3). The thickness and width of end-moraine ridges are approximated as random within defined bounds that are controlled by presumed sediment supply to the ice margin, the mobility of the ice front, and the ice-stream pattern. Time-dependent grids specify the lithology at the base of ice. An *interglacial–postglacial deposition module* forecasts thickness of debris accumulated between and after Weichselian erosion episodes, when additional automated time-slice modules estimate the possible thickness of interglacial sediments. This module was calibrated against Holocene offshore and onshore data. Figure 3.3 shows the pattern of sediment accumulation. Any sediment pockets could be individually resolved, depending on input grid resolution.

The results of this kind of analysis are illustrated in a corridor that runs from the northern Gulf of Bothnia across Finland and the Gulf of Finland into the Russian Plain in Fig. 3.4. The northern shore of the Gulf of Finland marks the approximate northern border of the Baltic–White Sea lowland – the area that contains the most erodable material that was particularly affected by glacial erosion. Figure 3.4a shows how we believe this transect looked at the end of the Tertiary before it was affected by any continental glaciations. Figure 3.4b shows the erosion that was accomplished in the first glacial cycles, showing the situation just after one

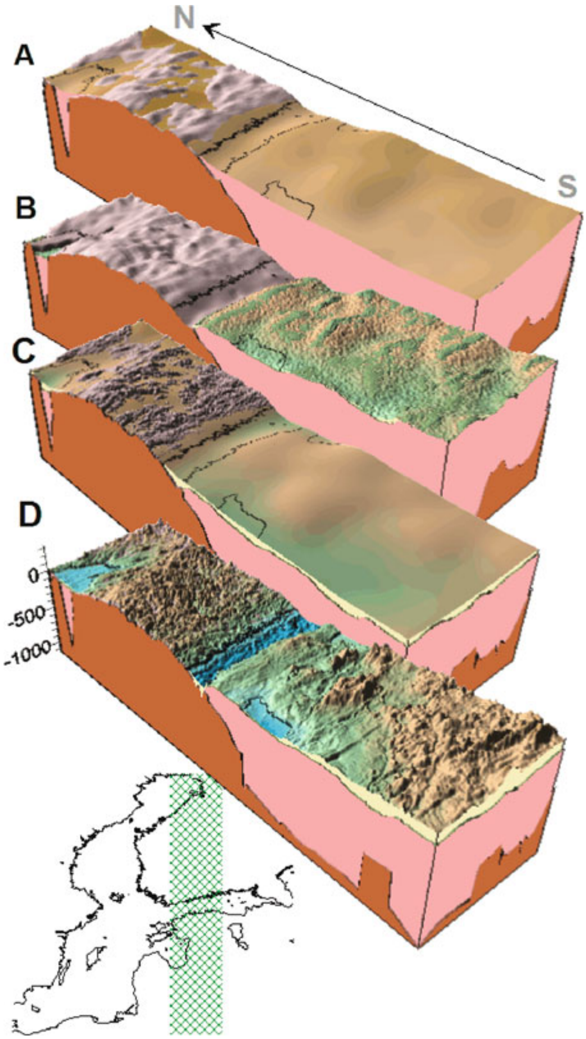


Fig. 3.4 3D slices showing simplified principal development of the Baltic region: **a** preglacial stage; **b** bedrock erosion of first major glacial impact; **c** later preglacial stage, illustrating input of interglacial sedimentation; **d** present. Color indications: basement – *dark red*, cover – *pink*, Quaternary cover – *yellow*

of these early cycles. The surface is rough and sculpted, and significant material has been completely removed from the Baltic–White Sea lowland area particularly. Figure 3.4c shows the situation at the end of the interglacial period that followed Fig. 3.4b. A layer of interglacial sediment has been laid over the rough lowland surface, and as a result the surface is smoothed in numerous areas. Finally, Fig. 3.4d shows the present situation that reflects intensive glacial erosion of mostly glacial and interglacial sediments, with resulting cumulative effects of all the Quaternary glacial cycles.

3.4 Results and Discussion

The analysis of sediment redistribution in the Baltic area using the methods sketched above is clearly a complex task, and to a considerable degree the validity of the methods used must be assessed by how geologically reasonable the product is perceived to be. The results of our analysis are described below, first in the areas peripheral to the Baltic–White Sea lowland and then in the lowland itself.

The Baltic–White Sea lowland (lowland for short) exhibits a regional first-order bedform that was to a significant degree created by strong diverse and multiphase glacial and fluvioglacial erosion of pliable sedimentary rocks covering the slope of the Baltic (Fennoscandian) shield (Amantov 1992, 1995). Its approximate shape is shown in Fig. 3.5. The lowland can be traced from the Baltic Proper with Gulf of Finland to the lakes Ladoga and Onega and then to the White Sea. It seems to have formed during rapid erosional lowering of wide Tertiary plains with the progressive removal of saprolites and less stable sediments. Basement features such as the sub-Cambrian or sub-Upper Vendian peneplains were exhumed around the present margin of the shield (Amantov 1995).

A narrow zone of eroded post-Late Vendian cover and Riphean–Jotnian formations is traced by the deepest indentations of the bedrock surface where hundreds of meters of unmetamorphosed rocks had been eroded. The deepest parts of the lowland usually coincide with areas where the sedimentary cover is truncated or, more rarely, with zones where the most friable sedimentary units thin. Major aquifers are often involved in the detachment of huge masses of cover. For example, the Gdov aquifer at the base of Late Vendian cover probably facilitated stripping along zones of disintegrated sandstone cementation and in areas with deep dissection by tunnel valleys or glacial hollows.

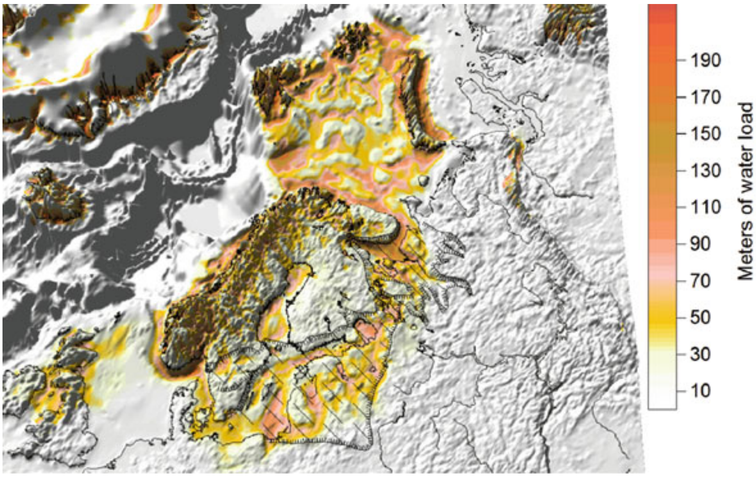


Fig. 3.5 Weischselian net erosion recalculated in meters of water load using averaged rock density. Baltic–White Sea erosion lowland is marked by slash pattern with half-ticks outline

The sub-Late Vendian or sub-Cambrian (basement) peneplain is an important reference for erosion up to where it is covered by sediments that are not penetrated by glacial erosion (Amantov 1992). It can be reconstructed on the adjacent shield area by interpolating preserved fragments of exposed peneplain under sedimentary cover and connecting summit highs of the Archean–Early Proterozoic crystalline bedrock. The slope of the stripped and slightly dissected peneplain presents one flange of the lowland onshore slope of parts of Sweden and Finland (Lidmar-Bergström 1992). The exhumed surface usually has shallow dip relative to its cover and is not significantly affected by faults. This contrasts strongly with the rugged (30–80 m) bedrock topography on crystalline rocks on the periphery of glaciated areas. There the topography is controlled by crystalline rock properties and structural peculiarities, like faults and fracture zones.

Bedrock depressions are often localized in the more erodable formations. They are separated by minor asymmetric basement highs whose steeper side faces the shield. The shallowest and narrowest lowland of this kind occurs in the Lake Onega–Vetryany Poyas region and on the Karelic peninsula, where elevation of the bedrock roof is 20–40 mbsl (meters below sea level), and locally overdeepened troughs with erosible lithology or structures can extend to 300–400 mbsl. On average the basement lies 50–200 mbsl. The depth to basement gradually increases from 55 mbsl in the eastern part of the Gulf of Finland to more than 200 mbsl in the Central Baltic Proper, where a paleo ice stream could have been located in the Gotland Deep.

Smoothed onshore scarps and slopes often bound the lowland. They are considered to be products of selective glacial denudation. Scarps and slopes usually face the shield, and their outline roughly corresponds to the outline of the ice at a particular glacial stage. The bounding is not distinct in areas where bedrock seems to be worn down and smoothed by ice streams. The more resistant strata control the plains between scarps and slopes. Evidence of their origin is provided by escarpments that can be traced in overdeepened locations like the >100 m scarps in the zone of maximal erosion in northern Lake Ladoga. These scarps can be seen to be localized by Riphean gabbroic sills that penetrated the sedimentary sequence (Amantov 1992, Amantov et al. 1995).

A significant percentage of the glacial erosion occurs in negative structures filled by more erodable, usually Riphean–Jotnian sequences. Examples are the Landsort, Åland, and Lake Ladoga deeps where the bedrock surface has been overdeepened by hundreds of meters (Amantov 1995). Rare thick Quaternary remnants in protected positions indicate the decisive role of first glaciations in excavating the troughs (Amantov 1992) and suggest that the subsequent glaciations mainly removed Quaternary sediments left by their predecessors and affected the bedrock surface in only a minor fashion. As a result, in zones of deepest erosion the tills now present belong mainly to the last glaciation and are overlain by the late-postglacial mantle. The last glacier could have removed 20–50 m of rocks of different density over wide zones of maximum erosion. Locally, in narrow overdeepenings, hollows, and glacial valleys, this figure increases to 70–90 m.

Tills, fluvioglacial, and other relevant sediments cover the peripheral accumulation belt. Late Pleistocene–Holocene uncompacted sediments starting with varved clays cover the entire Baltic Sea floor. Local sediment transfer is common. Numerous local overdeepenings of the late-glacial surface rapidly accumulated sediments immediately after glacial retreat. As a result, a thick (tens of meters over wide areas) veneer of sediments has been deposited. The impact on sediment loading is less than might be expected, however, because these postglacial clays are relatively uncompacted and have low density.

The central area of the Gulf of Bothnia is not a zone of low erosion as often expected from its position in the central zone of maximum ice accumulation. Lower erosion is expected in the northeastern part of the Bothnian Bay with adjacent onshore areas. In the western offshore part of the Bothnian Sea area, along the Swedish coast, erosion could be of the same order as in the Baltic Proper–White Sea lowland. According to our time-slice computations, much of the erosion occurred in the Early-Mid Weichselian, when ice marginal fluctuations occurred around the modern northwestern coast of the Bothnian Sea. Erosion was also strong during the piedmont phase and during glacial retreat.

In some ways the Swedish coast of the Bothnian Sea is comparable with the northwestern rim of the Lake Ladoga basin and other areas where unmetamorphosed Riphean–Jotnian sediments subcrop (Amantov 1992, Amantov et al. 1995, 1996). These areas are zones of deep glacial erosion. A west–east seismic profile from the Sundsvaal area (Axberg 1980, fig. 18) is similar to profiles crossing the coastal slope of the Northern Ladoga basin. Trends of bedrock topography are similar; even comparable scarps are observed in the Bothnian Sea in connection with resistant dolerite intrusions, but here they rise 25–30 m above the bedrock surface instead of 60–100 m in Lake Ladoga. The most intensive erosion resulted in the large negative relief form that is today the Hörnösand Deep. The present day bottom depths here range between 150 and 260 m, and the bedrock topography is slightly deeper. Such values are similar to those in the deepest northern part of the Lake Ladoga basin and to deeps in the steep coastal zone. Climate, the duration of ice activity, and ice streams can account for lateral changes of bedrock overdeepening along the contact zone between the crystalline rocks and the Riphean–Jotnian sediments.

Topographic similarities are connected with geological ones. In the authors' opinion, the deeps have been formed by the selective erosion of Riphean–Jotnian sandstones that fill tectonic basins. In the Hörnösand area erosion-resistant Ordovician limestones, which armor the bedrock surface to the south, are absent, thinning out at the southern slope. Here the erosion of the Cambrian–Ordovician platform produced a composite 100–120 m scarp-like slope that faces to the north. It is similar in form, magnitude, and lithology to the Cambrian–Ordovician slopes and escarpments in the Baltic Proper. The axial part of the Hörnösand Deep has sublongitudinal strike, joining to the south with a 100 m deep buried tunnel valley called the Aranda Rift (Winterhalter 1972). At some time-slices, an ice stream is expected southeast of the Hörnösand Deep and further toward the south, following an elongated bedform with depths between 110 and 160 m below sea level.

Locally, especially around the northern slope, Quaternary deposits up to 100–150 m thick occur in the Hörnösand Deep. A distinct acoustic appearance (Axberg 1980) may indicate that they belong to different glacial and interglacial events and are remnants that survived in shadow position, as in the Lake Ladoga basin. If so, this supports a scenario of excavating and shaping the major bedforms by the first glaciations, with subsequent oscillation between sedimentation and “cleaning out” of outlet zones.

In spite of the presence of pliable presumably Lower Cambrian–Upper Vendian sedimentary formations, the erosion of the northern part of the Baltic, the Bothnian Bay, is mild to moderate. This is supported by both bedrock topography and the pattern of glacial accumulation. The bedrock surface is rarely deeper than 130–170 m below sea level, and somewhat steeper along the Swedish coast. The southwestern area seems to have eroded, especially to the southeast from Shellefteo, but the erosion is mild compared to the Hörnösand Deep area. In the northeastern part of the Bothnian Bay, the bedrock surface on the Riphean sediments is 50–120 m lower than the surrounding crystalline rocks in the coastal area of Finland southeast of Ouly, where the Riphean–Upper Vendian Muhos formation comprises the half-graben appendage of the major Riphean–Jotnian basin. The bedrock is overlain by 50–80 m Quaternary sediments (Tynni and Uutela 1984). Thus, the bedrock surface is relatively deepened, as is noted everywhere where Riphean sediments are surrounded by harder crystalline basement, but to a lesser degree. The Quaternary sequence suggests moderate erosion prior to Weichselian. The total Quaternary section attains great thickness, frequently 50–100 m, and pre-Weichselian till deposits may be expected in the southwestern parts of the basin (Floden et al. 1979). Survival of the thick and complicated Quaternary succession in the subbottom area is in agreement with onshore observations. In the continuation of the major lowland in the Nordkalott area, north of the Bothnian Bay, the cover is comprised of two or more till beds, Eemian sediments are common, and even Saalian and older deposits occur (Hamborg et al. 1986). The survival of these remnants is compatible with their location in a complicated zone of ice divide, where the flow of ice was slow and its direction complicated with a dominance to the southeast (Hirvas and Tynni 1976). The first glaciations significantly transformed the region, by strongly eroding pliable terrigenous formations, which, together with the consequent isostatic adjustment, separated central sedimentary outliers of the Bothnian Sea and Bothnian Bay from each other and from the sedimentary domain of the Baltic Sea Proper.

3.4.1 Sediment accumulation and mass balance

Sediments accumulated around the areas glaciated as shown in Fig. 3.3. This sediment mass must, of course, match the mass of material eroded, taking into account the redistribution of material over a wider area. Our analysis assures that this is the case, not only today but also for every increment of erosion that occurred over the entire Quaternary (e.g., all the glacial cycles, including the last). There is great uncertainty regarding how the erosion is distributed between the glacial cycles, but we make an attempt to apportion it in a reasonable fashion.

The history of ice sheet development is relatively well known for the last 25,000 years, but uncertainties of earlier ice sheet oscillations are an important factor in possible model variations. In spite of uncertainties of imprecise estimations of erosion and accumulation rates in different areas, time-scale reconstructions provide a good picture of the regional loading–unloading cycles.

Our modeling also assumes variability of erosion and accumulation rates in time and space. For the Baltic area the largest short-term erosion rates are expected in the case when sediments are incorporated into the ice or pushed in front of glacier on initial advances in areas where intensive interglacial accumulation created unconsolidated extra-soft beds. Even on relatively hard argillaceous Late Vendian clays in eastern Gulf of Finland, the zone of very intensive dislocation has a normal thickness of 4–8 m with common thick slabs in overlaying tills. Increasing erosion rates during rapid deglaciation are related to highly dynamic ice masses, fluvioglacial processes, and outbursts from glacial lakes.

Modeling shows that the deepest sedimentary bedrock erosion is related to soft formations in depressions, i.e., graben-like structures, proximal to ice-flow contact zones between rocks of highly contrasting erodability. In such cases, hard abrasive material comes to the ice–bedrock contact zone, while the contact zone usually forms a relatively steep slope, possibly providing rotational flow with a sufficient supply of fresh firm abrasive. Major aquifers may serve as an additional factor in bedrock removed by other mechanisms.

Knowledge of bedrock topography and measure of its overdeepening and lowering from reconstructions of older geomorphic facets serve as important validation steps in the determination of the erosion magnitude. However, it cannot be used to judge erosion rates. In many cases, glacially shaped topography, with elongated basins alternating with conformal ridges and riegels, produced multiple local depocenters for interglacial (postglacial) sedimentation, partly being inherited. For such basins, erosion and later sedimentation could be compared with a pendulum, when the nature "masked its wounds." Local zones of deep erosion appeared as zones of thick sedimentation with maximum rates immediately after glacial retreat, but roles reversed again on the next advance. The initial glaciation(s) affected the bedrock, but later ones eroded glacial and interglacial deposits over wide areas (Fig. 3.4). We think that further development of joint simulation of different processes could be productive, in spite of the multiple assumptions and imperfection of our current simple tools.

The load redistribution caused by erosion and sedimentation is compensated isostatically. To assess this, sediment thicknesses must be converted to mass. Where the conditions are submarine, the load is the equivalent buoyant load. Whether on land or submerged, the porosity of the sediments must be taken into account. The algorithms we have designed take these matters into account. Figure 3.6 (right) shows the isostatic uplift and subsidence pattern that would be produced by the sediment redistribution that we estimate occurred over the last glacial cycle. Full isostatic equilibrium is assumed and the load is filtered by a lithosphere of flexural rigidity 10^{23} Nm (effective elastic thickness of 20 km) (Fjeldskaar et al. 2000). The modeling shows that the isostatic response to erosion and sediment loading (Fig. 3.6 (right)) is significant compared to that caused by deglaciation and sea level changes. The rise of sea level caused ca. 40 m of hydro isostatic subsidence under

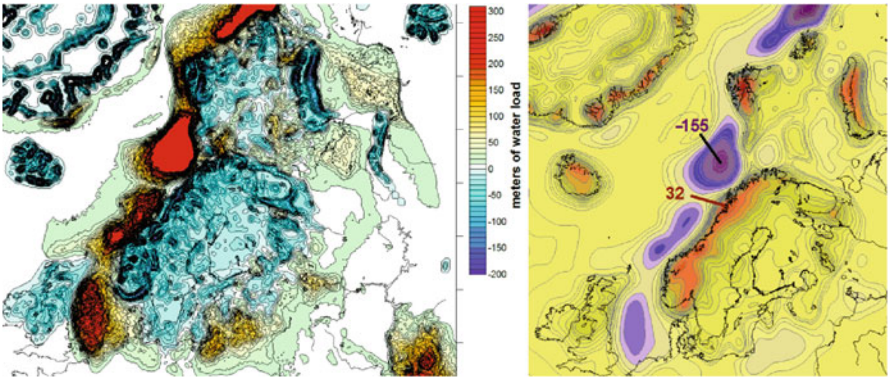


Fig. 3.6 Weischselian erosion and accumulation redistribution load recalculated in meters of water load using averaged rock density (*left*) and its total isostatic uplift–subsidence effect in meters (*right*)

the ocean load. Figure 3.6 (right) shows that the sediment loading of marine areas can cause isostatic subsidence five times greater than the loading by glacial meltwater. The uplift associated with erosion is smaller (<10% of the glacial isostasy) for the Baltic area, but for some areas of coastal Norway it could be a significant part of the observed postglacial uplift.

3.5 Conclusions

Although it is possible that bedrock erosion was evenly distributed between all the glacial cycles, most of the modification of the bedrock surface and shaping major overdeepened troughs was probably accomplished by the first glaciations of the Quaternary. Younger glaciations mainly removed sediments deposited by previous glacial cycles, reducing the thickness of the Quaternary succession and locally incising the bedrock surface. The isostatic effect of the glacial erosion and sedimentation significantly impact the total postglacial rebound. Subsidence in submarine areas adjacent to the continental glaciers can be much larger than that induced by the postglacial rise in sea level. Isostatic uplift caused by erosion is minor for the Baltic area, but could be a significant part of the observed postglacial uplift in coastal areas of Norway.

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