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Deformation of an early Preboreal deposit at Nykvarn (SE Sweden) as a result of the bulldozing effect of a grounding iceberg

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Abstract

An early Preboreal glaciomarine deposit near Nykvarn (SE Sweden) consists, from bottom to top, of a submarine outwash fan, a full (glacio)marine deposit, and a near-coast marine unit of poorly sorted ice-rafted debris. The top part of the succession shows soft-sediment deformations that have the same characteristics as material moved forward by a bulldozer. These deformation structures are interpreted as being due to grounding of an eastward-moving iceberg. This caused detachment of the lower part of the succession, in the way that sand is pushed forward by a bulldozer. The iceberg most likely calved off the retreating ice front (which extended into sea) due to a sudden subglacial discharge of meltwater (jökulhlaup), which event seems related to the breakthrough of an ice dam that could no longer withstand the pressure of a growing water volume produced by increased ice melting.

Keywords: Glaciomarine environment; Grounding; Iceberg; Soft-sediment deformation

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1. Introduction

The recession of the ice cap in central Sweden at the end of the last glaciation and the gradual change of the Baltic Ice Lake into the Yoldia Sea, which took place during the late Younger Dryas and the early Holocene, have been discussed for many years (by, among others, De Geer, 1896; Lundqvist, 1990; Björck, 1995). The sediments deposited in the Yoldia Sea contain dropstones, indicating that the sea was occasionally reached by icebergs.

The genesis of iceberg-related structures was discussed already by a number of researchers (among them Eyles et al., 1997). Nevertheless, no data on the impact of a sudden subglacial discharge as a trigger mechanism for iceberg release and on the induced deformation structures caused by the grounding of these icebergs have been published, as far as we are aware.

It seems probable that such a sudden discharge caused enhanced calving of the nearby Scandinavian ice sheet and, hence, the formation of an iceberg that grounded near Nykvarn (SE Sweden). This hypothesis is based on the fact that simultaneously with the grounding of an iceberg at Nykvarn a huge jökulhlaup occurred at Pålamalm. This latter site is at the same latitude as the Hummlemora gravel pit of the site in Nykvarn (Fig. 1), and supposedly the two locations shared an ice-front position for some time (Mokhtari Fard, 2003). Such catastrophic events must exert a

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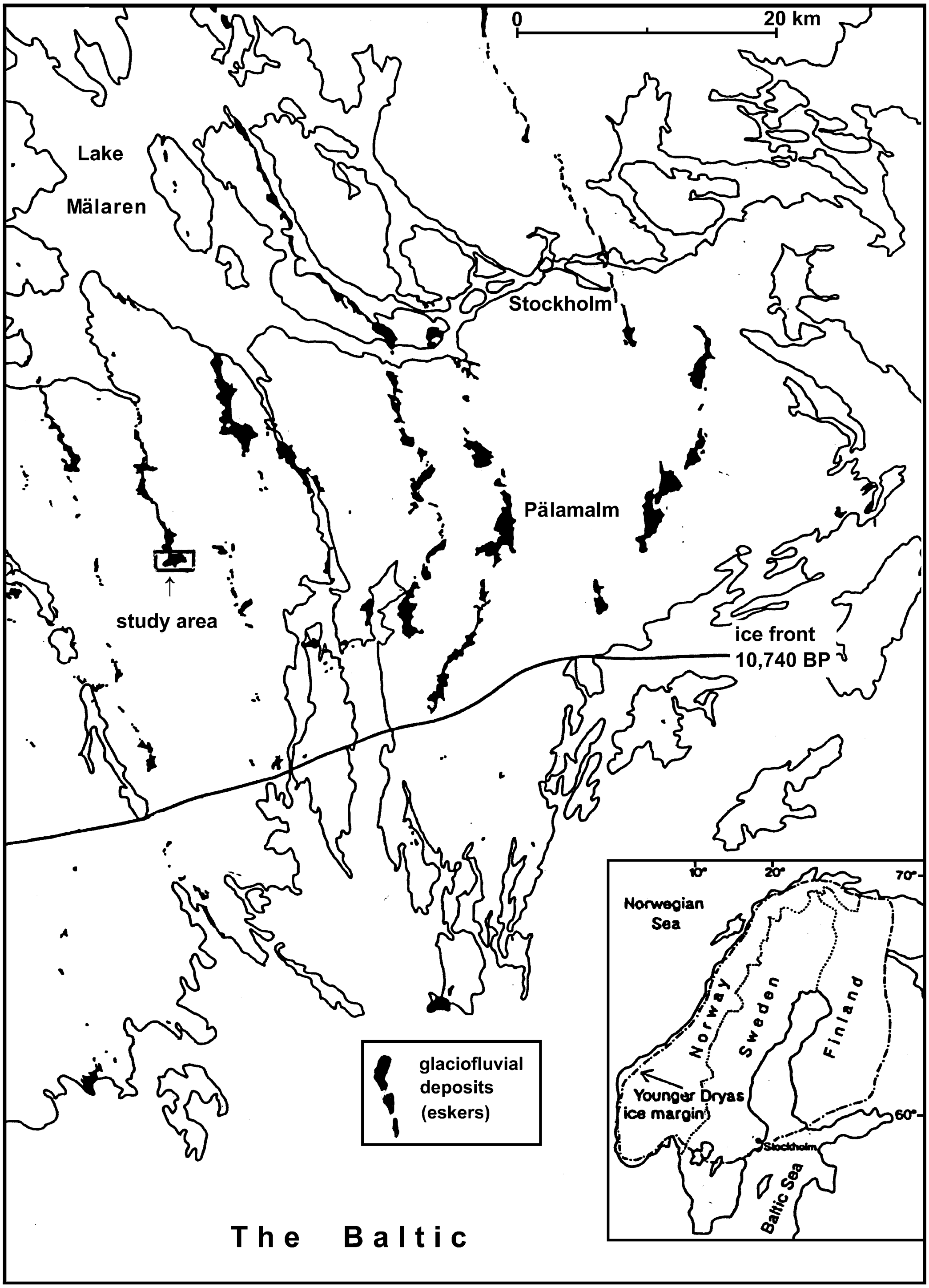

Fig. 1. Distribution of eskers in southeast Sweden, with location of the study area. The line marked '10,740 BP' indicates the extent of the ice during the final drainage of the Baltic Ice Lake and the opening of the Yoldia Sea Basin (modified after Brunnberg, 1995).

large pressure on the ice cap and facilitate sliding away of huge ice masses, and thus probably contributed to regional calving of the ice sheet.

The amelioration of the climate apparently favoured the formation of jökulhlaups which, in turn, triggered calving of the marine ice front. One of the icebergs that had been calved off apparently reached the glaciomarine outwash fan at Nykvarn which had developed shortly before in front of a subglacial, subaqueous conduit.

The present contribution focuses on the sedimentological processes at the outcrops exposed in the Hommelmora gravel pit in the Nykvarn site (which are reconstructed at the basis of the sedimentological characteristics and the facies transitions that occur, in combination with data from the literature about the paleogeographic development of the area during the time that these deposits were formed and deformed), and provides evidence for the deformational impact of one or more grounding icebergs on the glaciomarine succession that had previously formed at the site.

1.1. Geological setting

The distribution of glaciofluvial deposits in SE Sweden (Fig. 1) depends on the response of the Scandinavian ice sheet to the temperature rise during the late Younger Dryas interstadial (Lundqvist, 1990), because glaciofluvial deposition was controlled by the pressure/discharge relationship at the ice/bed interface. The glacial deposits in the Nykvarn area were formed just afterwards, in the early Preboreal (approximately 10,000 years ago), and are part of the Middle Swedish Ice-Marginal Zone (cf. Lundqvist, 1987); they overlie Precambrian gneisses of the Scandinavian Shield (Björnbom, 1981). Four zones representing stagnation phases of the retreating land-ice mass were identified in the region by Persson (1983). The study area, which is considered to represent an interlobate area between tongues of the retreating ice (Lundqvist, 1987), is located in Persson's (1983) zone 4, which is characterised by relatively thick accumulations of glaciofluvial deposits. These beds formed, in front of the rapidly retreating ice sheet, from huge volumes of meltwater that were set free during the Pleistocene/Holocene climatic amelioration (Brunnberg, 1995).

Most of these glaciofluvial deposits must have formed in subglacial tunnels; the majority might be characterised as tunnel-mouth deposits (cf. Brodzikowski and Van Loon, 1991). These elongated sedimentary bodies are expressed morphologically as eskers. Four such eskers have been identified in the area (Björnbom, 1981). The Hummelmora gravel pit (Fig. 2), where the present study was carried out, is located in the southernmost part of one of the eskers, the Turingestråket (Björnbom, 1981). It is located near the small town of Nykvarn, about 80 km southwest of Stockholm.

All eskers in the Nykvarn area show a distinct N–S trending orientation but the Hummelmora deposit has a locally deviating (E–W) direction. This E–W oriented deposit is about 1800 m long and 800

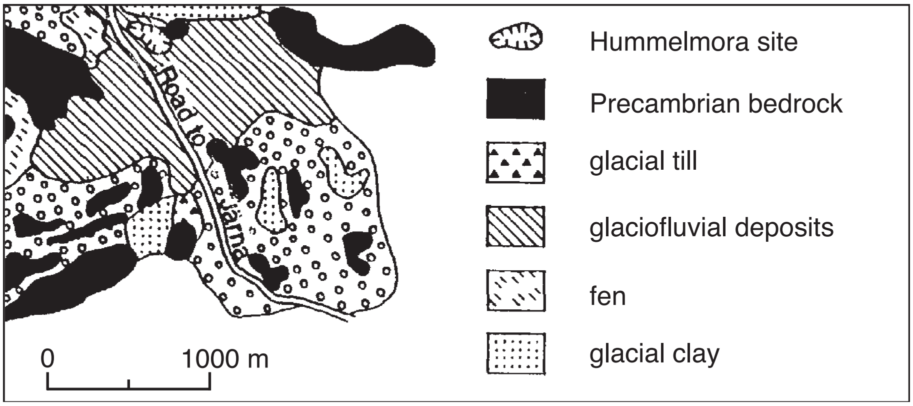

Fig. 2. Geological setting of the Hummelmora gravel pit.

m wide. Its maximum thickness is about 10 m. Considering the highest level reached by the Pleistocene coastline in the area (approximately 150 m above the present-day sea level) and the present-day elevation of the sediments, the deposit under investigation must have formed at a depth of about 100 m below sea level. The regional paleocurrent direction during the last glaciation was, as indicated by glacial striae, towards the southeast (Björnbom, 1981). Part of the sediments show large deformation structures of uncommon type. It is these deformation structures of which the origin is discussed in the present contribution.

1.2. Methods of investigation

The genesis of the remarkable deformation structures that were found in a gravel pit at Nykvarn could not be reconstructed easily. For this reason it was decided to decipher first their geological context; the sediments are described in the following sections. They were exposed in two opposing walls of the gravel pit (Fig. 3). Both the northern and southern walls are approximately 250 m long and up to 8 m high. Lithological profiles were measured and analysed in 12 sections (numbered 1–12); a generalised vertical log is presented in Fig. 4. Paleocurrents were measured on the basis of the preferred orientations of the long axes of gravel-size clasts and of cross-bedding in sand.

2. Facies 1

Facies 1 is the lowermost facies and is exposed in the lower part of both walls in the gravel pit (Fig. 3). It is the coarsest of the three facies distinguished, forming a 2–8 m thick succession of graded gravels (subfacies 1-a) and stratified gravel and sand beds (subfacies 1-b). The individual beds, which are up to about 25 cm thick, are laterally continuous for several meters and have non-erosional, planar to subplanar boundaries.

Paleocurrents indicate a roughly southward transport direction, the same as indicated by the striations on the Precambrian bedrock.

2.1. Subfacies 1-a

2.1.1. Description of subfacies 1-a

This subfacies represents the basal part of the succession studied; the contact with underlying deposits (or bedrock) is not exposed. The upper boundary is

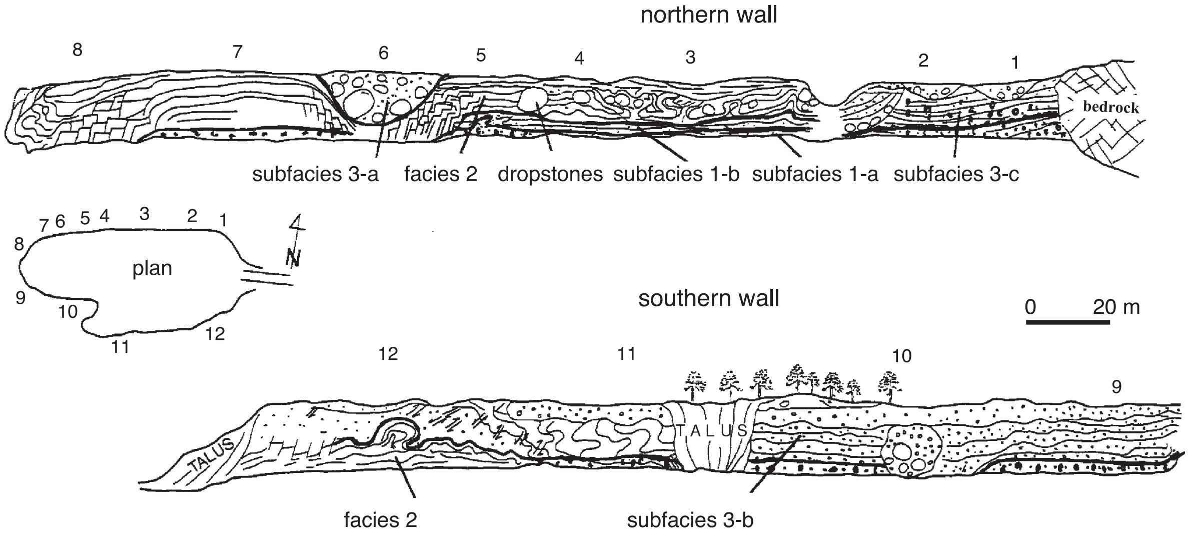

Fig. 3. Schematic drawing of the N and S walls of the Hummelmora gravel pit (the E and W walls are not exposed). The numbers 1 through 12 refer to logged sections. In the northern wall, which shows intense deformations, facies 1 is covered by facies 2; it contains dropstones in its middle part. The uppermost sediments are formed by the diamicton of subfacies 3-a. The S wall consists of deformed and faulted facies 2, covered by subfacies 3-b.

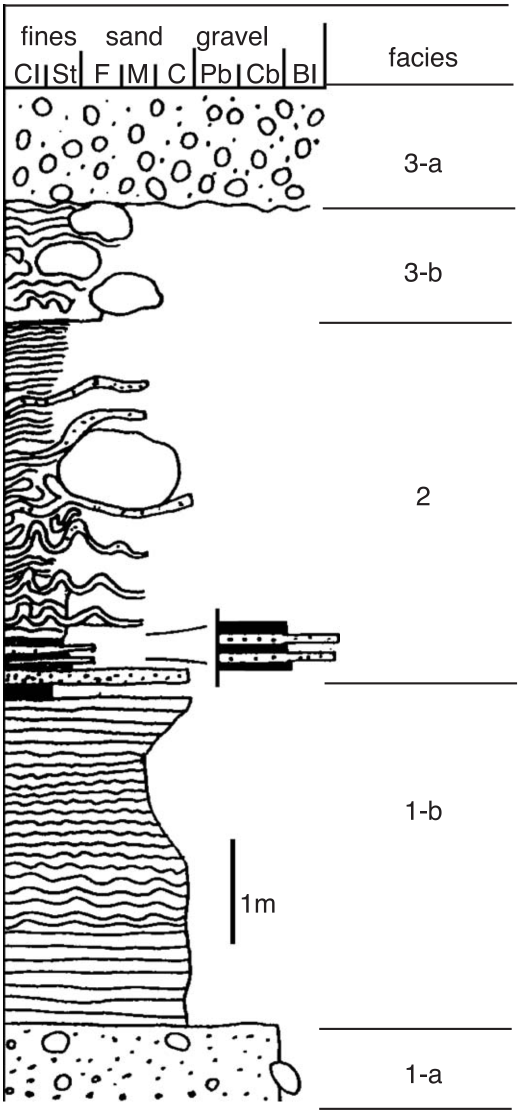

Fig. 4. Composite log of the succession in the Hummelmora gravel pit.

sharp and non-erosive. The subfacies consists mainly of tabular, horizontal to subhorizontal, matrix-supported sandy gravels. They are moderately to poorly sorted and show parallel stratification, frequently also an overall normal grading (Fig. 5). Bed thicknesses vary from 0.5 to 6 m; individual beds may change their thickness laterally. To the west, the beds become finer-grained and then pass laterally gradually into the planar pebbly gravel beds of subfacies 1-b.

The clasts in subfacies 1-a are pebbles and cobbles with a maximum size of about 10 cm. Some outsized clasts are found locally in the uppermost beds, where some beds are also characterised by horizontally aligned clasts and by an upward increase in matrix content. The matrix consists of medium to coarse sand.

In a small exposure in Section 9, the sediments of this subfacies are slightly overturned (Fig. 6) and more massive (crudely bedded) than in other sections. In addition, a collapse structure with high-angle normal faults is present between Sections 5 and 6, below the contact with subfacies 3-a (see Fig. 3).

2.1.2. Interpretation of subfacies 1-a

The coarse character and poor sorting of subfacies 1-a suggest deposition from high-energy currents with a relatively high sediment load. In particular, the horizontally bedded sandy gravels must have formed under the upper plane-bed conditions of the upper flow regime. In some sections, where the graded succession passes vertically gradually into subfacies 1-b, the transition reflects a gradual drop in energy level. The non-erosional character of the contact between subfacies 1-a and 1-b indicates a gradual change in energy, probably resulting from a change in discharge.

Collapse structures are common phenomena in coarse-grained materials that have been deposited on top of a buried ice mass (dead-ice structures). Small, high-angle normal faults record collapse after melting of the buried ice (cf. Rust and Romanelli, 1975). The disturbed sediments of subfacies 1-a are interpreted to be a slump structure, formed on a sloping sedimentary surface developed after melting of a buried ice block.

2.2. Subfacies 1-b

2.2.1. Description of subfacies 1-b

Subfacies 1-b consists of planar beds (up to half a meter thick) of fining-upward sandy gravel, alternating with beds of pebbly sand. The lower boundary, with subfacies 1-a, is gradual (Fig. 7). The matrix of the pebbly layers consists of coarse to very coarse sand. The frequency of clasts decreases upward. The sands show commonly very fine lamination.

This subfacies is present in the western part of the site (Fig. 3, Sections 5–8) and laterally pinches out to

Fig. 5. Graded gravel of subfacies 1-a in Section 1. On top of the section, subfacies 1-b, the parallel-laminated sand with beds of matrix-supported gravels towards the base, is also visible.

the east. It is faulted in Section 7, tilted in Section 8, overturned in Sections 9–12 (Fig. 6), and deformed in Sections 11–12, where the thickness of the entire succession is only approximately 0.5 m; the true dip direction is 236° and the inclination of the overturned beds amounts to 62–86°.

2.2.2. Interpretation of subfacies 1-b

The stratified sandy gravels and the gravelly, finely laminated sands must have been deposited under somewhat lower energy conditions than prevailed during deposition of subfacies 1-a. The currents must, however, have been powerful enough to transport

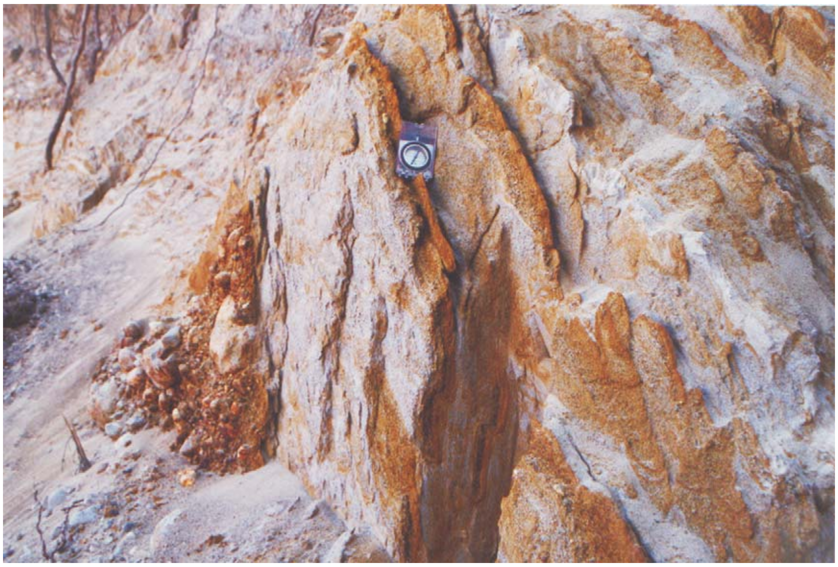

Fig. 6. Eastward view of Sections 9 and 10 in the S wall. The subvertical, overturned sediments of subfacies 1-a in the foreground dip 86° towards 200° (SSW). The position implies a structural relationship between these deformed deposits and subfacies 3-a.

pebbles and cobbles. The type of lamination indicates transport as a traction carpet (cf. Hiscott, 1994; Todd, 1996). A current velocity of 1 m/s is estimated to be necessary for the transport of clasts of about 15 cm diameter (Williams, 1983).

The horizontal bedding is attributed to the down-flow migration of sedimentary bed forms that were formed by interaction between eddies in the current and the bed (cf. Brennand, 1994). Subhorizontal lamination of sheet-like sediments is a well-known result of plane-bed conditions of the upper flow regime, which occur commonly during deposition of sand and pebbly gravel in subaquatic delta environments (Miall, 1996).

Well-defined, non-erosional vertical boundaries result from draping under waning-flow conditions. The common occurrence of gravel beds draped by thin sand beds suggests that en masse deposition of subaquatic gravelly debris flows was followed by deposition from turbidity currents (cf. Ghibaudo, 1992).

3. Facies 2

Facies 2 is the finest deposit of the site and is present over more than 50 m along the N wall of the gravel pit (Fig. 3, Section 12; Fig. 8). No subfacies are distinguished. The paleocurrents indicate roughly N–S transport, as in facies 1.

3.1. Description of facies 2

Facies 2 consists of alternating sets of fine sand, laminated clay, and clayey silt. The lower boundary of this facies (with both subfacies 1-a and 1-b) is sharp. The contact with the massive graded gravel of subfacies 1-a is visible in Sections 2–4 (Fig. 3) and is sharp and non-erosional.

The lower part of facies 2 consists of a sandy unit of some 30 cm (Fig. 4). The thicknesses of the individual sand layers decrease upward, whereas the thicknesses of the clayey silt beds show a reverse tendency. The thicknesses of the individual silty and clayey intercalations, which have a lateral extent up to

Fig. 7. Stratified sandy gravel of subfacies 3-a with intercalated gravelly sand of subfacies 1-b. A minor fault set forms the boundary of facies 3-a (Section 7). The scale bar is approximately 1 m.

Fig. 8. Strongly deformed glaciomarine facies 2 with large ice-supplied components (Section 3).

50 m, vary from a few millimeters to 5 cm. The alternations of fine sand and silt pass laterally into units of more clayey silt with minor sand beds of up to 0.5 cm thick. The sand beds are laterally continuous for several meters and have subplanar, non-erosional boundaries. In general, the sand and silt beds in this unit are well sorted and the only observed sedimentary structures—apart from lamination—are related to dropstones, which are common.

Deformation of the silty, clayey beds is observed where the thickness of the sand beds is at a minimum. Large-scale deformations (Fig. 9), including load structures, in facies 2 are associated with a number of outsized (up to 4 m) floating stones that disrupt the

Fig. 9. Strongly deformed silty/clayey glaciomarine deposit of facies 2, overlying stratified sandy gravel and sand of subfacies 1-b (lowermost layer).

bedding because their diameters exceed the thickness of the host strata. The floating stones push down—and sometimes penetrate—the underlying layers (Fig. 10). Silty and clayey beds in particular are strongly deformed in the direct vicinity of the stones, which are draped by overlying strata. In the non-deformed parts of the beds, the majority of the long axes of the clasts are oriented either parallel or perpendicular to the bedding planes.

3.2. Interpretation of facies 2

The presence of alternating sand and silt beds indicates minor, short-term fluctuations in the energy of the current that supplied these particles. A gradual overall decrease in energy is inferred on the basis of the vertical increase in the thickness of the silty/clayey sets and the upward decrease in thickness of the sand beds. The distinct separation of sandy and silty sediments suggests that these laminae resulted from alternating currents. The coarser sandy particles were probably transported by underflows, possibly in the form of small-scale turbidity currents; the silty and clayey layers, which have sharp upper and lower boundaries with the sands, represent settling of either the ‘tails’ of these turbidity currents or otherwise suspended material (for instance, supplied by overflow/interflow currents) during periods of almost stagnant water. The resulting deposits are fairly similar to those described by Ó Cofaigh et al. (2001) for turbid meltwater plumes that may have released icebergs into the sea.

The lateral discontinuity of this facies, in particular along the N wall, suggests that deposition was controlled by the basin topography. The Precambrian gneiss substratum is known from other locations to have a highly irregular surface, so that the sedimentary conditions must have changed drastically from place to place. The deepest parts of the sedimentary surface were presumably filled up first, whereas the highest parts remained void of fresh sediments. The pronounced gneissic ‘highs’ thus must be responsible for the depositional style that made facies 2 laterally discontinuous.

The isolated position of some large blocks, in combination with their distinct impact on the underlying strata, suggests that they are dropstones, supplied by floating icebergs. Their scattered occurrence in specific layers points to a restricted number of icebergs arriving. The flame, loadcast and water-escape structures in the deformed silty clay beds underneath the dropstones (cf. Lyså and Landvik, 1994; López-Gamundi and Martínez, 2000) record the effects of the suddenly increased pore-water pressure following the impact of the dropstones (cf. Knudsen and Marren, 2002; Phillips et al., 2002). It is noteworthy in this context that Ó Cofaigh and Dowdeswell (2001) attribute some glaciomarine laminated fine-grained sedi-

Fig. 10. The largest dropstone of the site with impact-related deformation structures in sediments of facies 2 (Section 4).

ments to settling from suspension plumes probably during an iceberg calving event. Another possibility is that such sediments resulted from pulses of meltwater influxes with suspended particles.

The overall architecture and the geological context of this facies suggest outwash conditions in a shallow-marine environment. The succession is considered to represent mainly deposition by low-density turbidity currents under the influence of ongoing meltwater-supplied sedimentation on a submarine delta-like body in front of the ice, prograding in a marine basin.

4. Facies 3

Lithofacies 3 consists of diamictons. Three subfacies are distinguished. The most important subfacies (3-a) is found in the N wall only, as is subfacies 3-b, whereas subfacies 3-c is exposed only in the eastern part of the S wall (Fig. 3). A gradual change in the current direction towards SE–SSE occurs towards the top of the succession.

4.1. Subfacies 3-a

4.1.1. Description of subfacies 3-a

This facies, which is exposed in a trough 15 m wide and 6 m deep (Fig. 3, Section 6), consists of massive and crudely stratified, matrix-supported diamictons that show a wide range of particle sizes. The individual beds within the diamictons are discontinuous; the beds become thinner towards the margins of the trough. Gravel, pebbles and cobbles, with occasionally boulders up to 3 m in size, occur in a sandy/silty matrix. An elongated boulder, approximately 1 m long, appears to have slid downslope along the trough's margin, disturbing the sediment in front of it. Some of the pebble beds are also folded and convoluted. The larger boulders are most frequent in the lower part of this subfacies.

Subfacies 3-a rests disconformably upon the succession of alternating laminated sands and gravelly sands of facies 1 and 2 that are cut by a series of parallel, closely spaced, mainly reversed faults dipping at steep angles away from the centre of the trough. The axes of the faults run parallel to the axis of the trough. The diamicton itself is unaffected by the faults, and the faults die out at, or immediately below, its base. Fault displacement is usually less than 20 cm.

4.1.2. Interpretation of subfacies 3-a

The lithology of subfacies 3-a differs strongly from the underlying facies 2: particles that form the matrix are roughly of the same size as the relatively large clasts in the glaciomarine sediments of facies 2. A genetic relationship is therefore likely. The isolated position of the diamicton in the form of a trough-like body within otherwise fine-grained marine sediment, the downfolding of the sediment underneath, and the nature of the associated faults together point to the grounding and subsequent in situ decay of a debris-rich iceberg.

As the iceberg became blocked by the sediments at the seafloor, its momentum was transferred to the bed, causing down-warping and faulting of the underlying sediment by water expulsion (cf. Thomas and Connel, 1985). Both the faulting within the surrounding sections in facies 2, and the overturned position of the sediments of facies 2 in Section 9, are interpreted as consequences of the contact of the drifting—and subsequently grounding—iceberg with the bottom sediments (cf. Woodworth-Lynas and Guigné, 1991).

4.2. Subfacies 3-b

4.2.1. Description of subfacies 3-b

The lateral extent of this unit is limited to Section 11, where it is exposed as a 2.5 m thick and 10 m long body, overlying a deformed part of facies 2. The boundaries of subfacies 3-b are sharp.

The massive, matrix-supported gravel contains clasts of pebble to cobble size within a matrix of coarse sand. Strong deformation occurs in this subfacies, including recumbent and overturned folds.

4.2.2. Interpretation of subfacies 3-b

The presence of structures beneath subfacies 3-b indicative of heavy loading suggests a relatively rapid rate of deposition on a water-saturated, soft-sediment substratum (cf. Benn and Evans, 1996). The similarity of the particle size between this unit and the underlying facies 2 suggests an uninterrupted supply of material, from the same source.

The faults affecting facies 2 cannot be traced into subfacies 3-b since they are truncated at the contact. This shows that deposition of facies 3 was preceded by a phase of faulting and subsequent erosion of facies 2.

4.3. Subfacies 3-c

4.3.1. Description of subfacies 3-c

Subfacies 3-c is a matrix-supported diamicton consisting of massive gravel with small, dish-shaped clasts of pebble and cobble size in a clayey matrix. It overlies facies 1 in Sections 1 and 2, where it has only a small lateral extent, up to 7 m (Fig. 3).

The diamicton is covered by deformed clay and silt. The lower boundary, with facies 1, is sharp, irregular and erosional.

4.3.2. Interpretation of subfacies 3-c

The distinctly erosional boundary and the lithological difference with the coarse-grained facies 1 suggest that the diamicton was deposited after erosion had affected the silts and clays that had been deposited in a glaciomarine environment (the silts and clays had been deposited in calm water, primarily by settling from suspension in an ice-marginal part of the sea, probably protected by a barrier or being a bay with a narrow inlet).

De Geer (1896) has described a comparable diamicton in a comparable geological context and attributed it to the release of material from a melting grounded iceberg, on the basis of its differences with other types of diamictons. We agree with this view. The outsized clasts within the clayey matrix were most probably deposited in a marine environment. The marine character is highly likely on the basis of earlier paleogeographic studies (Brunnberg, 1995), and the absence of varves is another argument against lacustrine sedimentation.

5. Overview of the environmental development

The depositional history has been reconstructed by sedimentological analysis of the lithofacies. Three units are recognized; they are indicative of, respectively, a submarine outwash fan, a glaciomarine basin and a near-coastal area where large icebergs touched the bottom and became blocked, eventually melting away completely.

The lowermost sediments are part of a submarine outwash fan that developed in front of a subglacial conduit, formed beneath an ice-sheet margin grounded in the Yoldia Sea. The gravels of this outwash fan (subfacies 1-a), which differ from the overlying glaciomarine deposits (subfacies 1-b) by the predominance of large particles, were deposited by subaquatic gravity-flows, which are common in subaqueous outwash environments. The lateral variation in the thickness of the unit is the result of localized deposition of sediment gravity flows on the low-relief, subaquatic surface of the outwash fan.

The sediments of the subaqueous fan (facies 1) are overlain by massive and normally graded fine- to medium-grained sand and massive to laminated silts and clayey silts of facies 2. The sand beds are products of high- and low-density turbidity currents. The silt and clay laminae interbedded with fine sand were deposited from suspension and some have undergone subsequent traction transport. This unit is consistent with deposition under low-energy conditions in a glaciomarine environment (cf. Horton and Schmitt, 1996). This situation may reflect that the ice-sheet margin had further retreated, so that the sediments record conditions in front of a delta-like fan prograding into the Yoldia Sea.

The diamicton of subfacies 3-a is closely associated with deformation of the subjacent facies and of two smaller diamicton outcrops. The various deformation structures and the general context—dead-ice structures and dropstones in a marine environment—in combination with the overturned gravelly subfacies 1-a and 1-b adjacent to diamicton subfacies 3-a and 3-b, suggest that the Hummelmora deposits result from deposition on a sea bottom that was mechanically disrupted by at least one grounding iceberg (Fig. 11). The presence of overturned beds of subfacies 1-a in Sections 9 and 10—in the S wall of the gravel pit (Fig. 3)—suggests a horizontal push mechanism that can—within this geological context—be explained satisfactorily only by a grounding iceberg. The same mechanism may have induced the faulting observed in the gravelly facies 1-b of Section 11.

6. Grounding icebergs and their marks

Studies in the present-day polar areas show that icebergs interact with the seafloor in extensive areas of the Arctic and Antarctic (cf. Barnes and Lien, 1988; Dowdeswell and Murray, 1990; Dowdeswell and Forsberg, 1992; Syvitski et al., 1996; Dowdeswell et al.,

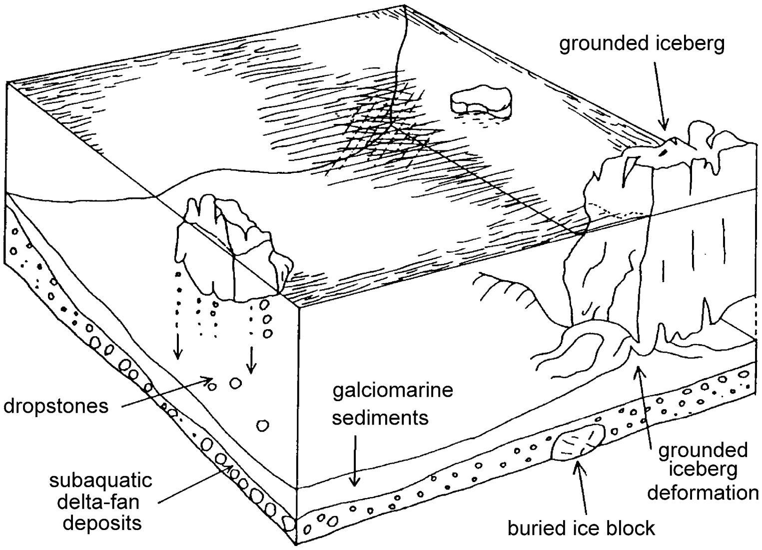

Fig. 11. Depositional model for the Hummelmora gravel deposit with a grounding iceberg deforming the soft-sediment bottom of the sea. Modified after Brodzikowski and Van Loon (1987).

2000). It is known that such icebergs scour the seabed and may produce significant debris layers, and that they did so also during the last ice age (see also Games, 2001), particularly after calving of huge ice masses (the so-called Heinrich events) (Dowdeswell et al., 1995; Ó Cofaigh et al., 2001). Yet, little is still known about the sedimentary characteristics of scours and related deposits, in spite of various approaches in the research, including numerical modelling (Yang and Poorooshasb, 1997) and seismic surveys (Lønne and Syvitski, 1997).

Icebergs must have reached the study area during the early Preboreal, possibly as a result of large-scale calving of the Scandinavian ice sheet. This could be related to temperature fluctuations at the time: evidence from North Atlantic deep-sea cores reveals a series of abrupt climatic changes, reflected in the cores as peaks in ice-rafted debris. Such a peak in ice-rafted debris—probably representing a large-scale ice-calving event—occurred approximately 10,300 years BP (Bond et al., 1997), more or less coeval with the formation of the deformation structures described here.

Stable-isotope investigation of marine sediments along the Swedish west coast suggests rapid deglaciation of the Scandinavian ice sheet during the early Preboreal. Several studies show a number of dramatic, climate-induced, ice-front fluctuations and drainage events for the early Preboreal Yoldia Sea in this area (Bodén et al., 1997). In the Stockholm area, marine cores through the Pleistocene/Holocene transition also contain ice-rafted debris (Andrén et al., 1999).

The similarity in the characteristics of the deformed sediments top those described in the literature from grounded icebergs, in combination with the nearby position of the Scandinavian ice front, suggests strongly that icebergs drifted into the study area during the early Preboreal. Scours from Holocene grounding icebergs have also been reported from elsewhere (among others, Gutierrez et al., 2003); they are also known from the Pleistocene (McHugh and Olson, 2002) and from pre-Pleistocene glaciations (Eyles et al., 1997; Ghienne, 2003). Vorren et al. (1983) introduced the term ‘iceberg turbate’ for seabed strata developed from the ploughing action of iceberg keels on the sea floor.

The interpretation of the depositional events yielding such glacial ‘markers’ is far from complete as there are no definitive sedimentological criteria for discrimination between glacially derived turbates and those generated by the keels of floating ice. Massive diamictons (also referred to as ‘waterlain till’) have been interpreted as formed either by debris release from melting of the lower part of ice bodies close to the grounding line beneath floating ice shelves or by grounded tidewater glaciers (cf. Pirrie et al., 1997). The rainout of clastic particles from melting icebergs is easily sedimentologically distinguishable from resedimentation by debris flows, among other criteria on the basis of fabric.

The floors of modern-day high-latitude continental shelves are subject to intense scouring by grounding icebergs (cf. Eyles et al., 1997), but also by shore ice (Dionne, 1998). Plough marks due to scour by grounded icebergs are common, especially near the sea margins where deformed sediments suggest that the shallowing bottom was pushed forward by grounding icebergs (cf. Bennett and Glasser, 1996). As the majority of modern icebergs are less than 50 m high and would not survive for more than two months after calving off (Ørheim, 1980), the presence of gravelly iceberg-supplied material suggests a depositional environment relatively close to the ice front.

7. Conclusions

The importance of subglacial hydrologic processes in the final disintegration of the Laurentide and Scandinavian ice sheets has been discussed elsewhere (Mokhtari Fard, 2000, 2003). The global impact of such events and the release of large volumes of freshwater into the North Atlantic ocean has been recently identified and the consequent iceberg discharges in this region in late Pleistocene is documented (Knutz et al., 2001).

The absence of laminations in the sediments above the assumed grounded-iceberg structure, along with the occurrence of the ice-rafted facies at the Hummelmora site, indicate a position close to the grounding line of the Scandinavian ice sheet. Comparison with the characteristics of modern grounded glaciers with rapidly calving ice fronts shows several similarities, among which the dense masses of debris. The fact that the dropstones at the Hummelmora site are restricted to a specific level (Fig. 10) implies, however, that the area was not frequently reached by icebergs, but a sudden arrival of a single iceberg—resulting in deposits like those at Hummelmora—was possible; recent facies assemblages comparable to those at Hummelmora have been described by Ashley (1995).

Combination of the above data makes it most probable that an iceberg reached the Hummelmora area, where it grounded and disturbed the underlying sediments (particularly those of facies 2). This is consistent with the fact that the area underwent rapid deglaciation during the early Preboreal (Brunnberg, 1995).

Considering the fact that the area is at the same latitude (and not far away) from another site (Pålamalm) where a giant jökulhlaup occurred at the same time, reaching the sea through a submarine tunnel, it is considered plausible that the jökulhlaup contributed to the calving of the ice front, thus possibly creating the iceberg that grounded at Nykvarn, thereby deforming the topmost sediments.

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References

Andrén, T., Björck, J., Johnsen, S., 1999. Correlation of Swedish glacial varves with the Greenland (GRIP) oxygen isotope record. J. Quat. Sci. 14, 361–371.

Ashley, G.M., 1995. Glaciolacustrine environments. In: Menzies, J. (Ed.), Modern Glacial Environments—Processes, Dynamics and Sediments. Butterworth, Oxford, UK, pp. 417–444.

Barnes, P.W., Lien, R., 1988. Icebergs rework shelf sediments to 500 m off Antarctica. Geology 16, 1130–1133.

Benn, D.I., Evans, D.J.A., 1996. The interpretation and classification of subglacially-deformed materials. Quat. Sci. Rev. 15, 23–52.

Bennett, M.R., Glasser, N.F., 1996. Glacial Geology: Ice Sheets and Landforms Wiley, Chichester. 364 pp.

Björck, S., 1995. A review of the history of the Baltic Sea, 13.0–8.0 ka BP. Quat. Int. 27, 19–40.

Björnbom, S., 1981. Beskrivning till jordartskartan Strängnäs SO. SGU Ae 39. 72 pp.

Bodén, P., Fairbanks, R.G., Wright, J.D., Burckle, L.H., 1997. High-resolution stable isotope records from southwest Sweden: the drainage of the Baltic Ice Lake and Younger Dryas ice margin oscillations. Paleoceanography 12, 39–49.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priorre, P., Cullen, H., Hajdas, I., Bonani, G., 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and Glacial climates. Science 287, 1257–1265.

Brennand, T.A., 1994. Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime. Sediment. Geol. 91, 9–55.

Brodzikowski, K., Van Loon, A.J., 1987. A systematic classification of glacial and periglacial environments, facies and deposits. Earth-Sci. Rev. 24, 297–381.

Brodzikowski, K., Van Loon, A.J., 1991. Glacigenic sedimentsDevelopments in Sedimentology, vol. 49. Elsevier, Amsterdam. 674 pp.

Brunnberg, L., 1995. Clay–varve chronology and deglaciation during the Younger Dryas and Preboreal in the easternmost part of the Middle Swedish Ice Marginal Zone. Quaternaria, vol. A2. Dept. Quat. Res., Stockholm Univ. 94 pp.

De Geer, G., 1896. Ändmoräner vid Venerns ostligaste bugt. Geol. Fören. Stockh. Förh. 18, 110–111.

Dionne, J.-C., 1998. Sedimentary structures made by shore ice in muddy tidal-flat deposits, St. Lawrence estuary, Québec. Sediment. Geol. 116, 261–274.

Dowdeswell, J.A., Forsberg, C.-F., 1992. The size and frequency of icebergs and bergy bits derived from tidewater glaciers in Kongsfjorden, northwest Spitsbergen. Polar Res. 11, 65–73.

Dowdeswell, J.A., Murray, T., 1990. Modelling rates of sedimentation from icebergs. In: Dowdeswell, J.A., Scourse, J.D. (Eds.), Glaciomarine Environments: Processes and Sediments. Spec. Publ.-Geol. Soc. Lond., vol. 53, pp. 121–137.

Dowdeswell, J.A., Maslin, M.A., Andrews, J.T., McCave, I.N., 1995. Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. Geology 23, 301–304.

Dowdeswell, J.A., Whittington, R.J., Jennings, A.E., Andrews, J.T., Mackensen, A., Marienfeld, P., 2000. An origin for laminated glacimarine sediments through sea-ice build-up and suppressed iceberg rafting. Sedimentology 47, 557–576.

Eyles, N., Eyles, Ch., Gostin, V.A., 1997. Iceberg rafting and scouring in the Early Permian Shoalhaven Group of New South Wales, Australia: evidence of Heinrich-like events? Palaeogeogr. Palaeoclimatol. Palaeoecol. 136, 1–17.

Games, K.P., 2001. Evidence of shallow gas above the Connemara oil accumulation, Block 26/28, Porcupine Basin. In: Shannon, P.M., Laughton, P.D.W., Corcoran, D.V. (Eds.), The Petroleum Exploration of Ireland's Offshore Basins. Spec. Publ.-Geol. Soc. Lond., vol. 188, pp. 361–374.

Ghibaudo, G., 1992. Subaqueous sediment gravity flow deposits: practical criteria for their field description and classification. Sedimentology 39, 423–454.

Ghienne, J.-F., 2003. Late Ordovician sedimentary environments, glacial cycles, and post-glacial transgression in the Taoudeni Basin, West Africa. Palaeogeogr. Palaeoclimatol. Palaeoecol. 189, 117–145.

Gutierrez, B.T., Uchupi, E., Driscoll, N.W., Aubrey, D.G., 2003. Relative sea-level rise and the development of valley-fill and shallow-water sequences in Nantucket Sound, Massachusetts. Mar. Geol. 193, 295–314.

Hiscott, R.N., 1994. Traction-carpet stratification in turbidites—fact or fiction? J. Sed. Res., A 64 (2), 204–208.

Horton, B.K., Schmitt, J.G., 1996. Sedimentology of a lacustrine fan-delta system, Miocene Horse Camp Formation, Navada, USA. Sedimentology 43, 133–155.

Knudsen, Ó., Marren, Ph.M., 2002. Sedimentation in a volcanically dammed valley, Brúarjökull, northeast Iceland. Quat. Sci. Rev. 21, 1677–1692.

Knutz, P.C., Austin, W.E.N., Jones, E.J.W., 2001. Millennial-scale depositional cycles related to British ice sheet variability and North Atlantic paleocirculation since 45 kyr BP, Barra Fan, UK margin. Paleoceanography 16, 53–64.

Lønne, I., Syvitski, J.P., 1997. Effects of the readvance of an ice margin on the seismic character of the underlying sediment. Mar. Geol. 143, 81–102.

López-Gamundi, O., Martínez, M., 2000. Evidence of glacial abrasion in the Calingasta–Uspallata and western Paganzo Basins, mid-Carboniferous of western Argentina. Palaeogeogr. Palaeoclimatol. Palaeoecol. 159, 145–165.

Lundqvist, J., 1987. Glaciodynamics of the Younger Dryas Marginal Zone in Scandinavia. Geogr. Ann. 69A, 305–319.

Lundqvist, J., 1990. The Younger Dryas event in Scandinavia. In: Lundqvist, J., Saarnisto, M. (Eds.), Termination of the Pleistocene—Excursion and Abstracts, Field Conference Norway–Sweden–Finland, May 9–16, 1990. Geol. Surv. Finl. Guide, vol. 31, pp. 5–90.

Lyså, A., Landvik, J.Y., 1994. Cyclic changes in the sedimentary environment during the last interglacial/glacial cycle; coastal Jameson Land, East Greenland. Palaeogeogr. Palaeoclimatol. Palaeoecol. 112, 143–156.

McHugh, C.M.G., Olson, H.C., 2002. Pleistocene chronology of continental margin sedimentation: new insights into traditional models, New Jersey. Mar. Geol. 186, 389–411.

Miall, A.D., 1996. The geology of fluvial deposits, sedimentary facies, basin analysis, and petroleum geology. Springer Verlag, Berlin, 582 pp.

Mokhtari Fard, A., 2000. The subglacial initiation process of the final drainage of Lake Agassiz. Geological Society of America, Annual Meeting. November 12–16, Reno, Nevada, USA.

Mokhtari Fard, A., 2003. Large dead-ice depressions in flat-topped eskers: evidence of a Preboreal jökulhlaup in the Stockholm area, SE Sweden. Glob. Planet. Change 35, 273–295.

Ó Cofaigh, C., Dowdeswell, J.A., 2001. Laminated sediments in glacimarine environments: diagnostic criteria for their interpretation. Quat. Sci. Rev. 20, 1411–1436.

Ó Cofaigh, C., Dowdeswell, J.A., Pudsey, C.J., 2001. Late Quaternary iceberg rafting along the Antarctic Peninsula continental rise in the Weddell and Scotia Seas. Quat. Res. 56, 308–321.

Ørheim, O., 1980. Physical characteristics and life expectancy of tabular Antarctic icebergs. Ann. Glaciol. 1, 11–18.

Persson, Ch., 1983. Glacial deposits and the central Swedish end moraine zone in eastern Sweden. In: Ehlers, J. (Ed.), Glacial Deposits in North-West Europe. Balkema, Rotterdam, pp. 131–139.

Phillips, E.R., Evans, D.J.A., Auton, C.A., 2002. Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. Sediment. Geol. 149, 157–182.

Pirrie, D., Crame, J.A., Riding, J.B., Butcher, A.R., Taylor, P.D., 1997. Miocene glaciomarine sedimentation in the northern Antarctic Peninsula region: the stratigraphy and sedimentology of the Hobbs Glacier Formation, James Ross Island. Geol. Mag. 136, 747–762.

Rust, B.R., Romanelli, R., 1975. Late Quaternary subaquatic outwash deposits near Ottawa, Canada. In: Jopling, A.V., McDonald, B.C. (Eds.), Glaciofluvial and Glaciolacustrine Sedimentation Soc. Econ. Paleont. Mineral. Spec., pp. 177–192.

Syvitski, J.P.M., Andrews, J.T., Dowdeswell, J.A., 1996. Sediment deposition in an iceberg-dominated glacimarine environment, East Greenland: basin fill implications. Glob. Planet. Change 12, 251–270.

Thomas, G.S.P., Connel, R.J., 1985. Iceberg drop, dump and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. J. Sediment. Petrol. 55, 243–249.

Todd, S.P., 1996. Sedimentary structures. In: Carling, P.A., Dawson, M.R. (Eds.), Fluvial Dynamics and Stratigraphy. Wiley, pp. 299–350.

Vorren, T.O., Hald, M., Edvardsen, M., Lind-Hansen, O.-W., 1983. Glacigenic sediments and sedimentary environments on continental shelves: general principles with a case study from the Norwegian shelf. In: Ehlers, J. (Ed.), Glacial Deposits in North-West Europe. Balkema, Rotterdam, pp. 61–73.

Williams, G.P., 1983. Paleohydrological methods and some examples from Swedish fluvial environments: 1. Cobble and boulder deposits. Geogr. Ann. 65, 227–243.

Woodworth-Lynas, C.M.T., Guigné, J.Y., 1991. Iceberg scours in the geological record: examples from glacial Lake Agassiz. In: Dowdeswell, J.A., Scourse, J.D. (Eds.), Glacimarine Environments: Processes and Sediments. Spec. Publ.-Geol. Soc. Lond., vol. 53, pp. 217–223.

Yang, Q.S., Poorooshasb, H.B., 1997. Numerical modeling of seabed ice scour. Comput. Geotech. 21, 1–20.