The ‘Captured Ice Shelf’ Hypothesis and its Applicability to the Weichselian Glaciation

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THE 'CAPTURED ICE SHELF' HYPOTHESIS AND ITS APPLICABILITY TO THE WEICHSELIAN GLACIATION

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ABSTRACT. The early and late stages of the Weichselian glaciation of the Baltic Sea proper and the Baltic countries is tentatively suggested as having been by means of a thin ice with very low basal friction, to a great extent floating on a number of water-filled cavities. The terms 'captured ice shelf' and 'captured lake' are introduced, suggesting that a floating ice is the norm, rather than an exception, in this concept. The captured ice shelf-hypothesis states, that a moving floating ice may, if and when the entire perimeter of the water body is reached by this floating ice (or 'captured ice shelf'), develop an 'ice rim' that acts as a hydrostatic seal, so that the water under the captured ice shelf cannot get out (it is a 'captured lake'). Meltwater inflow to the captured lake will lift the captured ice shelf, until the base of the latter reaches the level of the threshold, at which time the water will escape in a jökulhlaup. The water can escape to the next captured lake in a chain, or to the extra-marginal environment. Tunnel valleys and mid-sea channels with dead ends are suggested to be the product of these kinds of jökulhlaups.

Introduction

When interpreting marine geological field data from the Scanian east coast, Bornholm Gat, the present author came to the awkward conclusion that the Weichselian ice must have floated to the coast, much like an ice shelf (Erlingsson 1990, p. 121).

An ice shelf in that area seems highly unlikely, since the Bornholm Deep is too shallow ($\leqslant$100 m) for an ice shelf of the kind found along the Antarctic shores today. But instead of merely instinctively rejecting the idea of a floating ice, a glaciological explanation has been sought as to how floating ice could have existed in the Baltic depression.

In the following the hypothesis will be presented, some of its consequences will be mentioned, and these will be compared to some field data. In the next paper in the present issue ("A computer model..."), a numerical simulation to test the physical principle behind the hypothesis is reported.

This treatment does by no means profess to be a complete evaluation of the hypothesis. The purpose of publishing the hypothesis at this stage is to enable researchers to take it into account when doing original field work.

Field data indicating floating ice

There are some field data that has been interpreted as caused by a floating ice. Drozdowski (1988) reports a stratigraphy in the lower Wisla valley (northern Poland) that he interpreted as having been formed close to the grounding line of a floating ice.

Based on the study of sonographs and pinger profiles, along with dives, the present author interpreted the genesis of the glacial deposits in the Bornholm Gat off the southeastern coast of Skåne, southernmost Sweden (Erlingsson 1990). The interpretation was mainly based on the acoustic character and the geometry of the glacial deposits, which indicated a genesis below floating ice grounded at depths of ca 25 to 50 metres. Also, off Simrishamn there are a series of low ridges at -35 to -40 m (Erlingsson 1990, p. 69), which are interpreted as formed by floating ice, either at the base of an ice shelf or by post-glacial lake ice. Zilliacus (1989) interpreted similar forms in western Finland as having been formed where floating ice became grounded, calling them de Geer moraines.

Björck *et al.* (1990) reports that in the Bornholm Deep, below ca 58 m depth, there are two sequences of varved clay, separated by a thin layer of gyttja clay. The ^{14}C-age of the gyttja clay is >38 kBP (personal comm., Svante Björck 1990). This means that it predates the late Weichselian glaci-

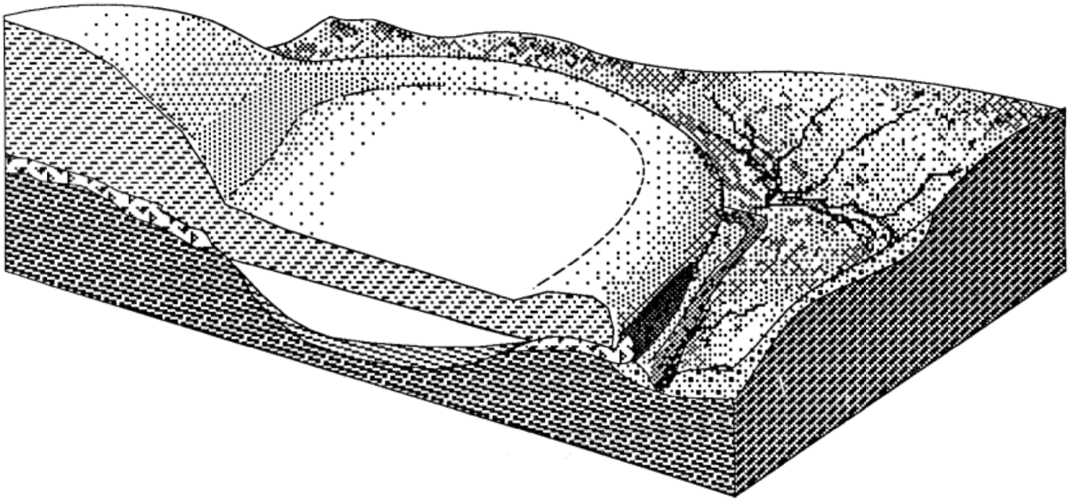

Fig. 1. A perspective plot of an ice sheet consisting of an inland ice and a captured ice shelf, below which the captured lake can be seen.

ation, which terminated roughly 13 kBP in that area. The fact that the older sediments are undistorted only below a specific depth may suggest that the ice floated above them; and the hiatus, that the sedimentation rate was extremely low beneath the ice.

On the nearby island of Bornholm, Lindström (1991, p. 30) suggests that the last ice advance (from easterly and southeasterly directions) might have been in the form of an ice shelf, in order to explain the absence of striae etc. on elevations above 100 m a.s.l. On the Österlen coast in Sweden, ca 40 km to the NW of Bornholm, there are several Baltic till units, the last of which was deposited by an ice coming from due south (Holst 1892, Vortisch 1972, 1977, Daniel 1986). The ice only reached one or a few kilometers inland, which means elevations of well below 100 m a.s.l.

Taken together, the above data suggests that the Weichselian ice never reached deeper than ca -58 m in the Bornholm Deep, and that the ice during one stage of the deglaciation did not reach over ca +100 m, nor under ca -58 m, in the Bornholm region. There was an isostatic gradient between these two places, which can be calculated roughly to 30 metres. The ice thickness can thus have been no more than ca 190 m.

When it comes to the water level, the best estimate may be had by taking 0.6 m/km to the NE as the isostatic gradient (cf. Erlingsson 1990, p. 90), and extrapolationg that from the Hanö Bay to Öresund. If Öresund acted as a threshold, then the water level at Bornholm would have been no more than a few metres over the present, allowing a maximal ice thickness of ca 60 m. If the ice on some occasion was 190 m thick, the water level on which the ice floated must have been ca 50 to 70 metres above the threshold in Öresund.

There seems to be a general agreement that the basal friction in the Baltic depression must have been low. As an illustration, the computer model of Boulton *et al.* (1985) showed that a low friction is required in order to explain the transport pattern of erratics. One common way to explain the low friction is by means of the presence of deformable subglacial sediments. The hypothesis presented in this paper offers an alternative explanation that may of course be used in conjunction with the hypothesis of deformable sediments where the ice was grounded.

The captured ice shelf hypothesis

Glaciological concept

Definitions: If an ice shelf completely covers a lake or a sea, and is grounded on all sides (Fig. 1), it becomes what will henceforth be called a *captured ice shelf* (a confined ice shelf is bordered by land on the sides but has a calving front). The waterbody that remains underneath will be referred to as a *captured lake* (Fig. 2).

The ice shelf may possibly originate as perennial sea or lake ice, as suggested in the Marine Ice Transgression Hypothesis (cf. Denton and Hughes

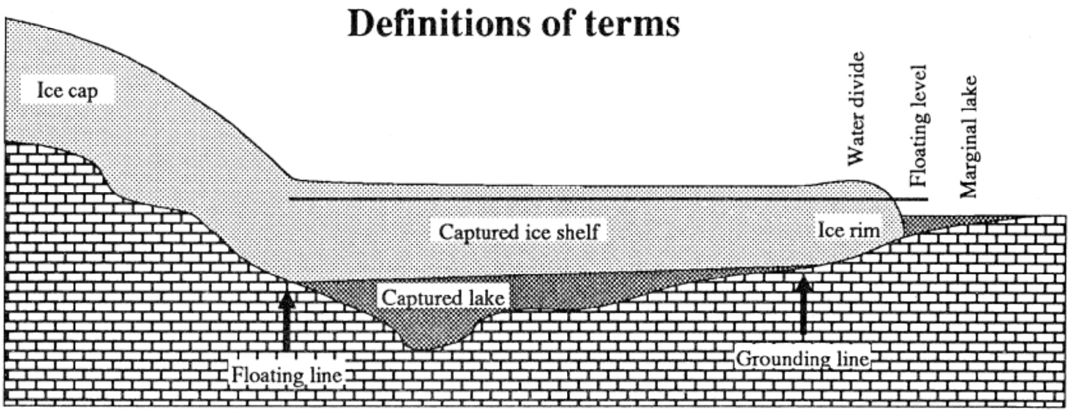

Fig. 2. Definitions of terms. The floating level is the level of the water surface if a hole is drilled through the ice. The captured lake ceiling equals the base of the captured ice shelf.

1981; Hughes 1987). The motion could be secondary, as glacier ice is fed into the lake. Actually, that would help to explain why the sediments in the Bornholm Deep are undisturbed, even though the water depth before the glaciation of that area was not more than some tens of metres. A 'regular-sized' ice shelf would not be able to advance over the area.

Where the captured ice shelf is grounded on an upslope, an ice rim will form. The reason for this is that the ice will be pushed up on the slope (Fig. 3). This ice rim thus forms a low 'ridge' surrounding the captured ice shelf. The marginal side of the ice rim may or may not display an ice wall—it does not affect the model as long as calving velocities do not exceed ice velocities.

The form of the ceiling of the captured lake is a mirror image of the ice surface, though ca ten times steeper (Björnsson 1988). Under the ice rim the higher elevation of the ice surface will force the equipotential line of the captured lake ceiling down under the ground level (Fig. 3). Water will be forced towards the captured lake from all points within the water divide of the ice rim (Björnsson

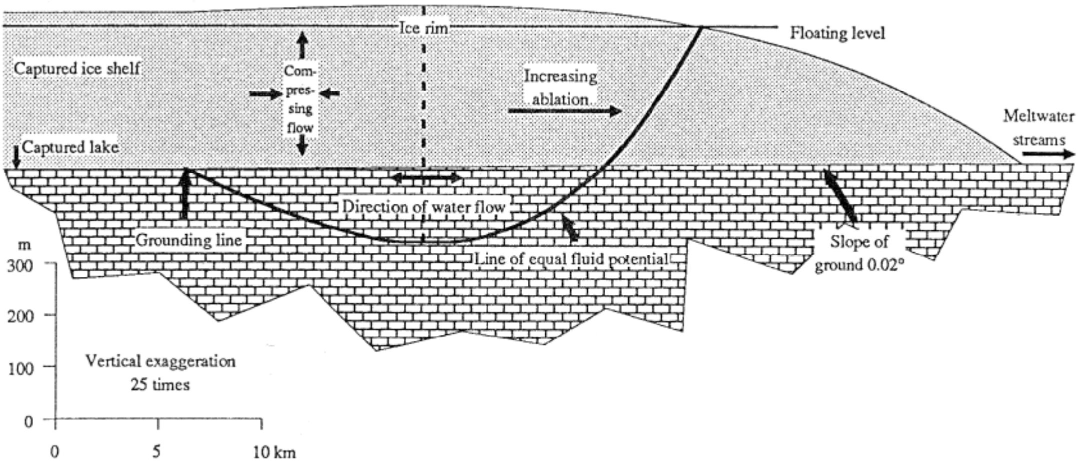

Fig. 3. The principle of the ice rim and the water divide. Note that the captured lake exists under the ice to the left of the grounding line. If the ground is impermeable the equipotential line will seal off the lake, so that the floating level inside the water divide is higher than the threshold elevation. The line of equal fluid potential below the ground level is hypothetical, since it assumes that the density of the ground equals that of ice. The ice rim is formed as a result of the ice being pushed up-slope faster than it melts. After a jökulhlaup compressing flow may also occur.

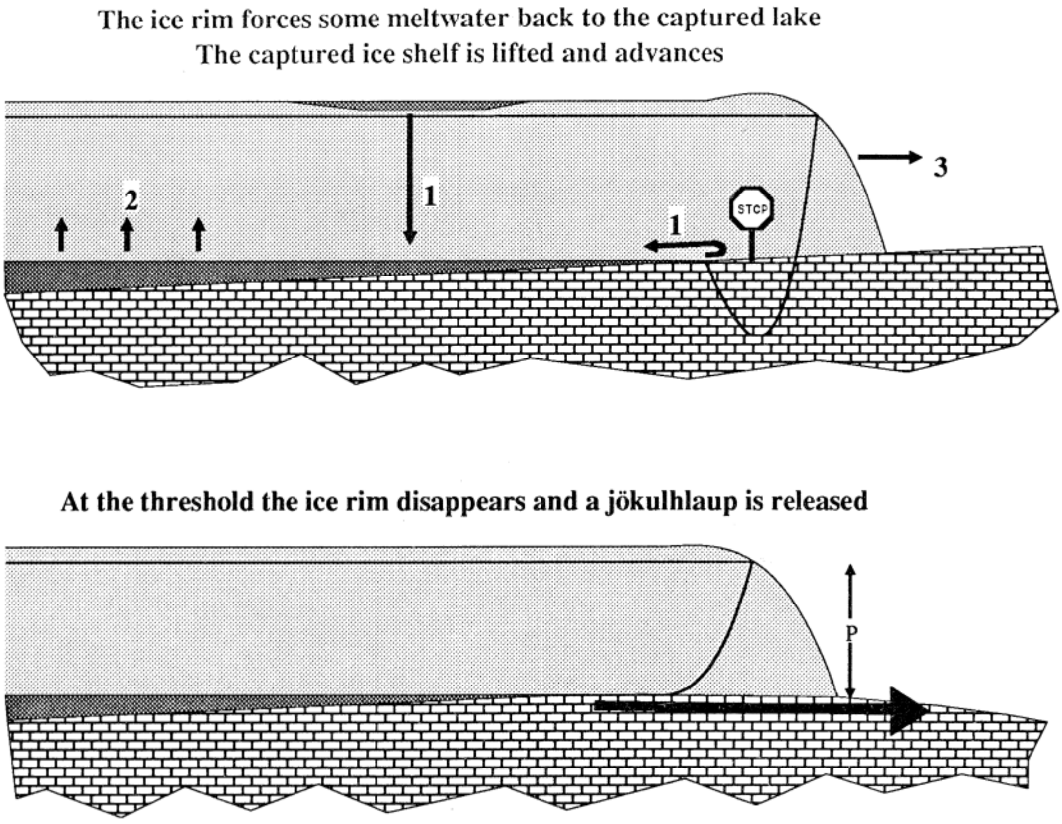

Fig. 4. The basic idea behind the captured ice shelf concept.

1988). Meltwater on top of the captured ice shelf will of course flow towards the lowest point, where lakes are formed. It can be argued, that these lakes might eventually be drained to the captured lake, through the ice shelf.

Water may be lost if refreezing occurs underneath, which will happen if low temperatures prevails. Melting underneath can only be caused by thermal heat from the earth's mantle. This is quite different from calving ice shelves, where the water circulation may bring in warm water under the ice, and cause rapid melting at its base.

The volume of the lake will increase if the inflow of meltwater exceeds refreezing underneath, and in that case the ice shelf will be lifted (Fig. 4). As a result, the bottom friction decreases and the ice shelf advances over new areas. This is the key factor in the life of a captured ice shelf. It means that for continued growth, the captured ice shelf must probably in part have a negative mass balance, or at least a supply of meltwater to the captured lake. A very cold climate on the other hand, would give a cold ice that would cause freezing underneath. It would lead to the formation of a cold-based inland ice instead.

When the ice front reaches a threshold locality water will start to flow out (Fig. 4). The outflowing water may enlarge the tunnel much more rapidly than the rate at which the ice overburden pressure can close it (Nye 1976, Björnsson 1988), in which case a jökulhlaup results. It can be argued, that the water may erode a tunnel valley in the substrate to a level given by the mirror image of the ice rim, ten times magnified.

The position of the threshold is given by a combination of the ground topography and the ice topography, where elevation differences in the latter are about ten times as important (Björnsson 1989). But since an ice shelf has a low gradient, the ground topography is definitly not unimportant.

If the position or level of the threshold is changed, either as a result of a change in mass balance, due to isostatic effects, due to damming by ice, or as a result of erosion, then the change of level in the captured lake will trigger an advance or retreat along the entire perimeter of the captured ice shelf. This implies that the changes in position of the captured ice shelf margin are, in the best situations, indirectly connected to climatic changes.

Some geomorphological implications

A large jökulhlaup may be considered a diagnostic feature for a captured ice shelf (in a non-volcanic region). As suggested above, the event might produce a tunnel valley. Furthermore, depending on the amount of water and debris transported, and on the extra-marginal topography, one might expect either an extra-marginal erosional channel and/or a sandur to be formed (cf. the morphology south of Vatnajökul). It remains to be examined whether the captured ice shelf hypothesis offers a credible explanation for the tunnel valleys and urström valleys found along, and inside, the ice margins of northern Europe.

After a jökulhlaup, the ice along the ice margin will stand harder on ground than immediately before. This may lead to the ice stagnating, and melting down as dead ice. This was also predicted in the model run (cf. next paper in the present issue). One end moraine may be left after each jökulhlaup, so that if the ice retreats by means of a number of jökulhlaups at successively lower levels, the same number of end moraines is expected. This hypothesis thus offers an explanation to the generation of many end moraines in a relatively short time period—and without any fluctuations in the climate whatsoever.

On the other hand, if the captured lake finds a new threshold radically lower than the previous one, then the resulting large jökulhlaup would give a wide zone of dead ice morphology. The difference compared to the above scenario is entirely caused by the topography, and not by the climate.

A captured ice shelf is suggested to be able to transport debris over long distances. Since there is no water exchange with a non-ice-covered ocean, the only sources of heat are the geothermal heat (which can give melting only in the range of a few centimetres per year), and the atmosphere, which will produce melting at the top and not at the base of the ice. Where the mean annual temperature is well below freezing and the ice is in contact with the bottom, the ice may pick up debris at its base as water freezes on to the ice. The debris may be transported a long distance with the relatively rapidly moving floating ice. As the thermal regime becomes warmer, melting will occure primarily at the top of the captured ice shelf. The debris will therefore be transported to the ice front, where it is likely to be deposited as an end moraine in a rather 'wet' environment, since the ice is likely to end with a cliff in shallow water, be it standing or running.

To sum up, compared to a traditional inland ice sheet, the debris is transported a long distance in a short time (i.e., from a freezing to a melting area); practically no part of it is deposited on the way, no lodgement till is produced, but 'glaciofluvium' may be formed under the ice, and almost all of the debris will be deposited close to the ice front in the presence of melt water. In the melting zone a debris cover may be formed on top of the floating ice; this may re-freeze and in connection with the deglaciation be deposited in the sedimentary silt and clay under the ice as clasts of melt-out till.

During deglaciation rapid disintegration is possible and likely when 'capturing' ceases. 'Meltout till' may be deposited on the grounded portion, grading laterally into a slightly sorted lacustrine or marine diamicton in deeper waters. This will be followed by the sedimentation of varved silt and clay. For a general summary on rafting in the glacimarine environment, refer to Gilbert (1990).

Isostatic implications

A captured ice shelf will give rise to some isostatic pressure when the floating level (cf. Fig. 2) is above the sea level, but it will give zero *differential* pressure, pressure differences being impossible in the free water body under the ice. An isostatic gradient could, however, be caused by the existence of several captured lakes in a succession, each at a lower level. The stepwise lower isostatic pressure would give an isostatic gradient on a regional scale, the fine details being averaged out by the crusts flexure radius (ca 180 km, cf. Walcott 1970).

A Weichselian glaciation scenario

This scenario is an attempt at applying the hypothesis to the Weichselian glaciation in the Baltic Sea proper, and examining how field data from the area can be explained by this hypothesis. Since the key area for the hypothesis is the threshold region, the emphasis will be on that.

The pleniglacial

The assumption in this scenario is that the Weichselian ice sheet developed a captured ice shelf in the Baltic Sea. The floating line could, at one point in time, have been parallel to the coast of eastern Sweden-Åland-western Finland.

The ice shelf would have moved parallel to the east coast of southern Sweden. That might explain the abundance of roches moutonnés in the archipelago: ice shelves are very effective when it comes to eroding small obstacles, whereas larger obstacles like Gotland will become covered by a protective ice cap, an ice rise.

Farther south, in the Kalmarsund area, Rudmark (1980, p. 28–31) summarizes the occurances of striae. There are striae from directions between N50°E and N65°W, the easterly ones (N25°E) in Örntorp older than the westerly ones. On the mountain Alkullaberget, farther east, diffuse striae in N10°W occur on the top, and very coarse striae in N25°E occur together with diffuse striae in N50°E on the steep eastern slope. The easterly striae might be caused by an ice shelf in the Baltic Sea, which was followed by an advance of the ice sheet on the Swedish mainland that created the westerly striae.

The inland ice sheet reached Skåne in southern Sweden around 21,000 BP (Lagerlund 1987). A thickening ice over the threshold region in Denmark must have forced the captured lake to find new thresholds farther south and away from the feeding inland ice in southern Sweden, i.e., over Germany. During deglaciation, the opposite may have occured. In the process, many tunnel valleys, end moraines, and till units would have been formed. The geomorphology of the landscape westwards from Oder is indeed one of tunnel valleys, abundant end moraines, and numerous till units. The latters are treated in, e.g., Kronborg *et al.* (1990).

The final threshold region

The deglaciation of Öresund is known to have been complex. In this scenario, as successively lower thresholds opened in the course of the deglaciation, the captured ice shelf could no longer float above southern Skåne, but increasingly followed the trend of the present coastline. This eventually led to an "ice stream with sharp limits mainly following Öresund" (Ringberg 1989, p. 331). The ice movement direction must have shifted gradually from the NE to the S, which is in agreement with field data (Lagerlund 1980).

The transgression in Kattegat at ca 13,500 BP (cf. Ringberg 1989) may have lifted the front of the ice shelf and allowed it to advance. The following regression may then have caused any ice shelf in Öresund to become (at least partially) grounded again. This could be the explanation of the ice that is usually known as the Low Baltic ice (cf. Lagerlund 1980). This ice deposited the sediments termed 'Lund Diamicton' by Malmberg Persson and Lagerlund (1990), and 'Malmö till' by Ringberg (1989). According to Ringberg (1989, p. 326, 334) the ice melted away as dead ice, during a regression down to ca 15 m a.s.l.

It can be noted that Lagerlund (1980) interpreted Lund Diamicton as a glacimarine sediment with load casts from ice, and not as a till. That interpretation was disputed by Ringberg (1989), who gave the following description (p. 333): "This ice moved rapidly, probably 0.5–1.0 km per year … ; it had a low profile and gives no indication of being a heavy load on the earth's crust". Ringberg did not find any example of the sedimented and later glacially tectonized deposits that Lagerlund found, but one find is enough to conclude that the ice was afloat for at least a short while (at Limhamn, by Malmö). The base of the ice must have been above the present sea level, and since the water level according to Ringberg (1989, Fig. 14) never was higher than ca +45 m, the ice can not have been thicker than ca 50 m. The field data of Lagerlund and Ringberg appear incompatible with a grounded ice, but the ice shelf hypothesis allows for a floating ice in Malmö at the same time as a grounded ice in, e.g., Lund (further 'down-glacier' but at a higher elevation). However, this does not indicate wether the ice shelf was captured or not.

Referring to Duphorn *et al.* (1979), Ringberg states that the duration of the Low Baltic readvance was 13,200–13,000 BP. Fairbanks (1989, Fig. 2) found that the world sea level rose by as much as 24 metres in the period 12,500 to 11,500 ^{14}C-years BP. The absolute-age estimate of this presumed melt-water pulse was 13,070 BP. It might have been this sea-level rise that caused the transgression in Kattegat and thus the Low Baltic readvance, if the world sea-level rose faster than the earth's crust in the area of Öresund. Since the magnitude of the latter is probably in the order of several metres per hundred years, the presumed melt-water pulse must have been very concentrated.

Disintegration of the ice shelf

Any captured ice shelf in the Baltic Sea must eventually have become 'un-captured', which would give the water off the ice front the opportunity of flowing in under the ice and cause basal melting. The effect would be a rapid disintegration, possibly salt water inflow, and a much increased sedimentation.

Field data from southern Baltic Sea suggests that the ice cover indeed disappeared rapidly: the gyttja clay mentioned in the introduction (dated to an infinite ^{14}C-age) corresponds to varve -120 of Antevs (1915). Consequently, varve -119 must be the late-glacial bottom varve. From a varve diagram published by Duphorn *et al.* (1981, Fig. 6), it is evident that this varve is synchronous over a large area (Christiansø, offshore Blekinge, Kristianstad plain). Björck *et al.* (1990, p. 274) mention that the varve -200 corresponds to 12,770 varve years BP, which would mean that the disappearance of the ice in the Hanö Bay occurred about the varve year 12,690 BP (approximately the same in ^{14}C years). Arpi and Utech (1967) found an increase in the salinity in the interstitial water in the varves above the gyttja clay layer, which they interpreted as a sign of marine influence. This suggests that the world ocean was in connection with the early Baltic Ice 'Lake' via the Danish Bælts and/or Öresund.

The captured ice shelf hypothesis offers an explanation to this rapid change of sedimentary environment: a bottom varve directly above a hiatus of ⩾24,000 years, and simultaneously over a large area.

Indicator clast composition

The second to last till in southwestern Skåne (Dalby Till; Lagerlund 1980) contains an increasing amount of Cretaceous chalk in the upper part, but only below a certain elevation (i.e., till on high elevations lacks the chalk). The last till in the area (Malmö till) also contains large ratios of chalk, and the till is constrained to elevations below 90 m a.s.l. (Ringberg 1989). Similarly, there is an absence of Cretaceous chalk in the till at Berlin (Böse 1990). That till was deposited during the main stage of the Weichselian glaciation.

The lack of chalk indicates absence of erosion in the Arkona Basin (Böse 1990, Fig. 8). She suggested the presence of immobile ice in the basin, over which other ice moved without picking up any material. The captured ice shelf hypothesis offers an alternative explanation. The present depth in the basin is ca 45 m. According to Erlingsson (1990), the base of the ice shelf was at ca 40 m depth off southeastern Skåne when the present glacial morphology was created, and it was never lower than ca 55 m b.s.l. Thus, the Arkona Basin would have been below the captured lake ceiling during the maximal glaciation (i.e., when the Dalby Till was deposited on the higher elevations in Skåne, and the ice margin was at Berlin).

Furthermore, whereas most tills in Skåne only contain minor amounts of Paleozoic limestone (found in the area of Öland–Gotland; cf. Lagerlund 1980), tills on very low elevations frequently contain large ratios of it (e.g. the oldest till on the south coast, Lagerlund 1980; the youngest till on the east coast, Åmark 1984; the oldest till in central Skåne, Ringberg 1988; the (only?) till in the Bornholm Deep, Björck *et al.* 1990; and the increasing content upwards in the Dalby Till on low elevations in SW Skåne, Ringberg 1989). This distribution pattern, with a strong correlation to elevation, is what one would expect from a captured ice shelf, that can erode the bottoms between Öland and Gotland only when the floating level is low.

The ice margin

The margin of a captured ice shelf can be predicted to follow the contour lines rather closely, which should then be reflected in the end moraines. Of course, minor fluctuations must be allowed for, since they may be caused by a number of factors, including post-glacial erosion. Therefore a first test was made on a continental scale, assuming that minor discrepancies would be 'evened out', and only the isostatic effects need to be addressed.

A digital map was used. Elevation data was taken from the UNEP-GRID data set "ETOPO-5", and the bathymetry from Swedish charts. The original resolution of the data was 5 minutes of longitude and latitude, but it was transformed to a conical projection with 5 km resolution (using the nearest neighbour interpolation method). The elevation resolution varies depending on the availability of map data for different regions, from 1 metre in most parts, to 5 or 10 and even 30 metres in the some parts of Russia (beyond the area of interest, though).

The synchronous position of the ice margin was digitized from a map in Andersen (1981; Fig. 1–1). The line for 14 kBP was selected. However, one

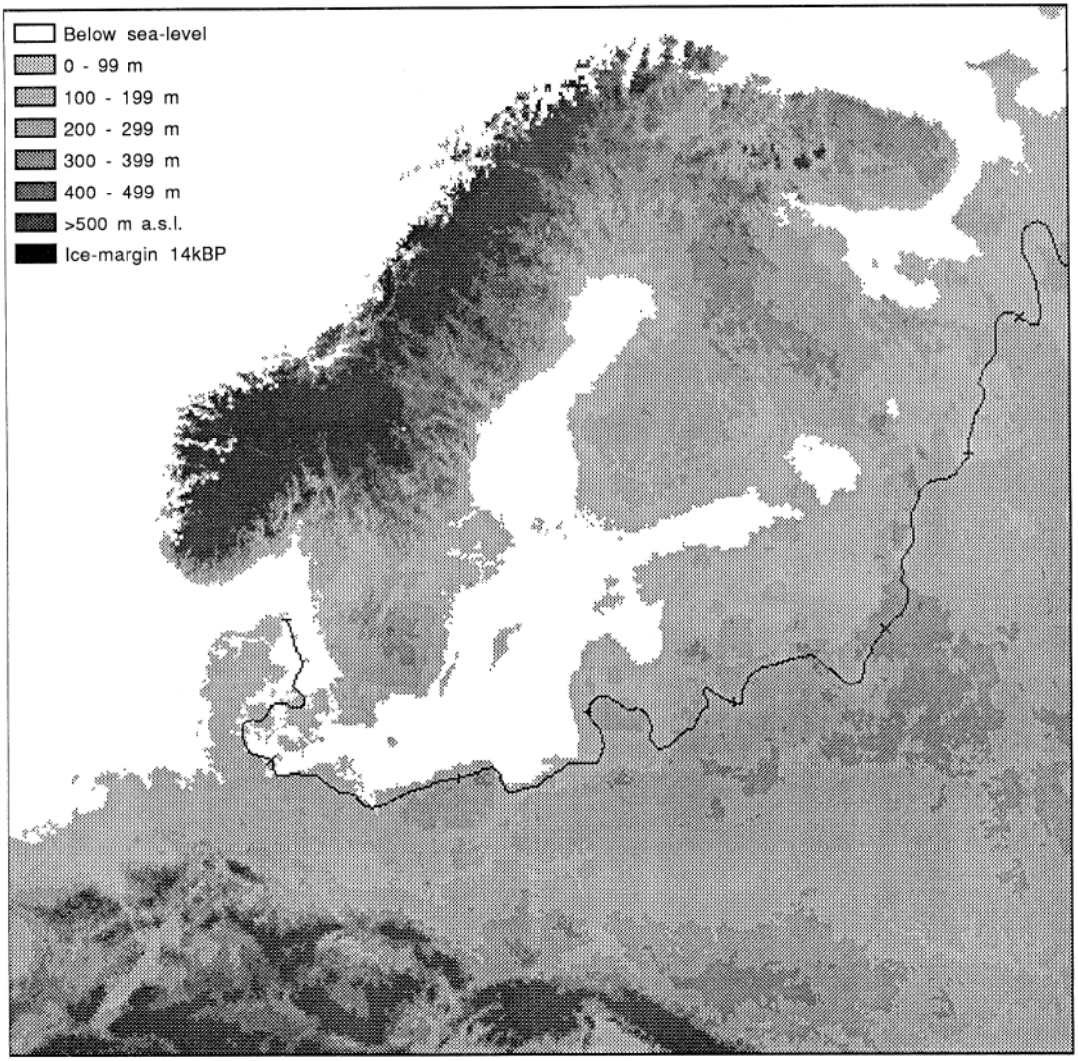

Fig. 5. The ice margin along which the profile in Fig. 6 is taken. The line corresponds to the 14,000 BP line from Andersen (1981, Fig. 1–1). A small mark is made at each 500 km along the ice margin, with Skagen as origin. The last mark stands for 3400 km, and is an exception.

has to bear in mind the possibility that the curve is not synchronous at all points. Nevertheless, it probably represents the most comprehensive analysis of data available so far.

The outermost pixel covered by the ice was selected (Fig. 5). The elevation of these pixels were plotted as a function of distance from Skagen (northernmost Denmark) *along the ice margin* (Fig. 6). They were grouped in 25 km distance classes, and an interpolation was performed using the software Cricket Graph (on a Macintosh). Thus, the line in Figure 6 indicates a smoothened height profile along the ice margin.

The methodics was checked in Lithuania. Figure 6 was compared with a recently published map over Lithuania in the scale 1:300,000 with 20 m contour interval. The precise position of the selected ice margin, "Northern Lithuanian Glaciation", was taken from "Atlas Litovskoi SSR" (1981), although it is perfectly obvious already from the topographical map where the end moraine is (almost 'lateral moraine' in this case). The

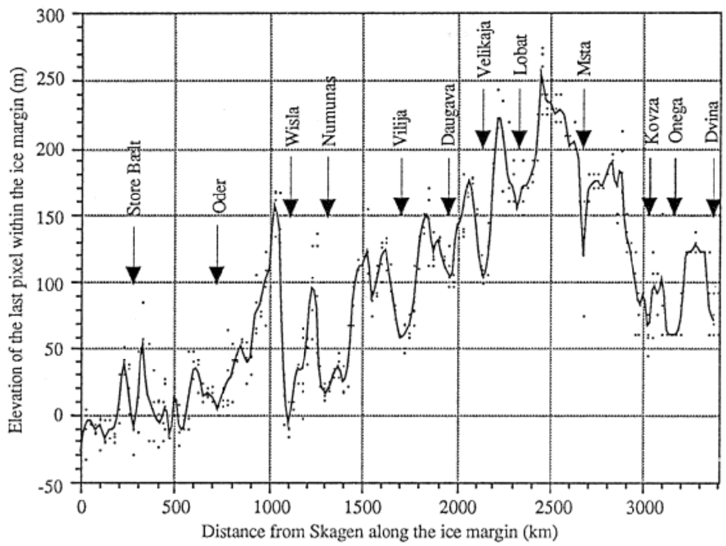

Fig. 6. The height profile of the ice margin in Fig. 5. The names of the major rivers etc. are given for reference.

end moraine has a very regular trend in Lithuania, sloping south by ca 0.4 m/km, and the variations are within ca 20 m except for locations where fluvial erosion has been active (the slope is discussed further in the following).

The conclusion is that Figure 6 contains quite a lot of artificial elevation variations. Disregarding them, the ice margin elevation between kilometers 500 and 2500 has a rising trend, interrupted by river valleys. The elevations in the latters are lower both due to postglacial fluvial erosion, and due to the differential isostatic land uplift (in the valleys the ice margin is farther away from the ice centre, and has thus suffered less postglacial isostatic rebound).

The only major exception from the regular trend is found in northern Poland, west of Wisla. That may be an artefact caused by the methodics. Since the terrain there is steep, a small horizontal error in the position of the ice margin (originating either from the original map, from the digitizing, or from the adjustment to a different coordinate system), would result in a large error in elevations.

Thus, one cannot exclude the possiblity that the end moraine was formed quasi-horizontally, although it is presently sloping from east to west.

The isostacy

The slope of the left leg of the U-shaped end moraine of the Northern Lithuanian Glaciation is as mentioned above roughly 0.4 m/km (due south). The slope is caused by two components: Ice-surface slope, and differential isostacy, the latter consisting of a combination of contemporary ice pressure and earlier ice pressure not yet rebound.

Calculations (cf. the Appendix) show that the isostatic gradient caused by a grounded ice with a surface slope of 0.4 m/km would have been in the order of 0.1 m/km (a slope of 0.4 m/km appears too low for the sides of a grounded ice, anyway). The gradient caused by a series of captured ice shelves could have been in the order of 0.2 m/km, which is more reasonable: From Svensson (1989, Fig. 188) it can be seen that the isostatic gradient just before the final drainage of the Baltic Ice Lake at ca 10,300 BP was ca 0.13–0.14 m/km in the nearby Gulf of Riga. Thus the value at 14 kBP must have been between 0.14 and 0.4 m/km, which is satisfied by the value 0.2 m/km that was calculated for a series of captured ice shelves.

As a comparison, the early deglaciation isostatic gradient for areas close to the ice dome has been found to be ca 0.6 m/km in the Hanö Bay (Erlingsson 1990, p. 89–91) and 0.59 m/km in Lake Vättern (Norrman 1964). Incidentally, the latter area, then a part of the Baltic Sea, also shows evidence of a floating ice in its southern parts before deglaciation (cf. Waldemarson 1986, e.g., p. 34.)

The conclusion to be drawn from these calculations is that neither a grounded inland ice nor a single captured ice shelf can explain the isostatic gradient. Instead an ice with a very low, but not nil, basal friction is needed. The low friction could be caused by a system of sub-glacial water-filled cavities, i.e., captured lakes. All of them need not have existed simultaneously. The up-glacier ones may have disappeared as the ice grew too thick during the pleniglacial. The only area where the present author suggests that the ice stayed afloat even during the pleniglacial, is in the Bornholm basin and possibly in the Gdansk and the Arkona basins, all just inside the southern ice margin. There may at that time have been at least two captured lakes, separated by a threshold in the Stolpe Furrow.

Returning to Lithuania, the end-moraine in western Lithuania that is correlated with the Northern Lithuanian Glaciation, has a markedly lower level than the latter: ca 40 m a.s.l. instead of the expected 80 m a.s.l. The Riga Bay has got a topography that suggests that a captured ice shelf there easily could have been separate from the one in the Gdansk Basin and/or Gotland Deep, which would explain the height difference.

The difference from a horizontal line is 200 m between km 500 and 2500 in Figure 6, corresponding to a slope of ca 0.125 m/km (the great circle distance is ca 1600 km). The slope could be interpreted as a stepwise change from east to west, and not a gradual one, the reason for this pattern being that there were several and separate captured ice shelves in existance at the same time. The difference in floating level between these captured lakes may have varied with time, since each one of them has its own threshold that regulates the jökulhlaup events. The isostatic effect would have been a gradual slope from east to west along the end moraine.

However, a part of the slope of 0.125 m/km is probably attributable to a remaining isostatic effect, since the ice margin in the east had retreated farther from the glacial maximum than it had in the west.

Böse (1990, p. 225) noted that the ice movement direction in northeastern Germany shifted from north to northeast and even to easterly directions during each Quaternary glaciation (directions inferred from origin of indicator clasts). This is consistent with the interpretation of a succession of captured ice lakes from east to west, since not only the water but also the ice will flow towards the lower elevation.

Summary

In many cases the hypothesis offers explanations that appear more straightforward than does any other hypothesis. This is true as regards the origin of tunnel valleys; the indicator clast composition in tills in the southwestern Baltic Sea area; the lower isostatic gradient on the eastern rim of the Baltic Sea than on the Scandinavian peninsula; and the rapid expansion and decay of the Scandinavian ice sheet.

Some field data indicate that captured ice shelves may have existed in the Baltic Sea area. It has been shown both in this paper and in the model run (cf. next paper in this issue), that a spatial and/ or temporal succession of several captured ice shelves is possible and probably necessary. Such a succession may have evolved both from the ice sheet towards the margin, and along the axis of the Baltic Sea from the Gulf of Finland to the Arkona Basin.

The hypothesis suggests a rapid expansion of the Weichselian ice over the Baltic Sea, followed by a relative standstill for thousands of years (although the thickness may have changed), again followed by a rapid decay. Before that rapid expansion, however, there must have been a long period of build-up of the Scandinavian ice sheet in northern Sweden and Norway.

The response time of a captured ice shelf is much shorter than that of a traditional inland ice sheet. The fact that it is thinner is only one reason. Another is that since water is accumulated in the captured lake, runoff of rain and melt-water is much reduced compared to a normal glacier. The depletion of water from the world oceans is for a captured ice shelf drainage area a function of the accumulation of ice *plus* water, and not of ice alone. Similarly, during deglaciation the return of water to the oceans is quicker since the water in the captured lake is already in liquid phase, and unlike ice it does not need energy to be melted.

More importantly, the captured ice shelf-glaciation is only in broad outlines controlled by climatic changes. Local topography at the margin, especially at the threshold, is the main control as regards the details of advances and retreats.

Consequently, compared to other hypotheses, this hypothesis implies a different coupling between climate, ice-margin position, and water storage in the great ice sheets. It also implies that climatic variations are virtually impossible to interpret in detail, if the interpretations are based only on ice margin fluctuations.

This model implies a somewhat smaller ice volume than a traditional ice sheet model. The difference for the Scandinavian ice sheet is, however, probably far less than 50%. In the present model, any sub-glacial lake water has to be included in the ice volume in such a comparison, which reduces the difference.

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Appendix: Calculation of the isostatic gradient

Assume first that the slope of the (lateral) 'end moraine' **E** is the sum of the differential isostatic upheaval **U** and the ice surface gradient **S**, that is, $E = U + S$. In the case of a plastic earth crust, $U \approx 0.33\,T$, where **T** is the ice thickness gradient ($0.9/2.7 \approx 0.33$, the density ratio between ice and the ground).

For *grounded ice* on a flat surface $S \approx T$. In the case of grounded ice $E \approx 0.33\,T + T$, or $1.33\,T$, which for a value of $E = 0.4$ m/km gives a value of $T = S \approx 0.3$ m/km and $U \approx$ *0.1 m/km*.

In a captured ice shelf $S \approx 0.1\,T$, and U will be zero within any single captured ice shelf. Since the flexural radius of the earth's crust is ca 180 km (cf. Walcott 1970) a regional isostatic effect will arise if there are *several captured ice lakes with different floating levels*, giving $U \approx 0.37\,F$, where **F** is the (regional) floating level gradient ($1/2.7 \approx 0.37$, the density ratio between water and the ground). Unfortunately, F and T are not related, so the equation for E cannot be solved directly: $E \approx 0.1\,T + 0.37\,F$. But if we know the value of T and E, we can calculate the value of F and thereby U. Taking the average width of the ice toungue to be 100 km, the ice thickness gradient T of an ice shelf in fresh water with a calving front, can be calculated to, roughly, 2 m/km (cf. Sanderson 1979). Using this value of T we can solve the equation: $F \approx 2.7(E - 0.1\,T)$, or $F \approx 0.54$ m/km, and $U \approx 0.37 * 0.54$. Both the ice surface gradient S and the differential isostatic gradient U calculates to *≈0.2 m/km*.

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