Quaternary Glaciations – Extent and Chronology
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Evidence of European ice sheet fluctuation during the last glacial cycle

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Abstract

Satellite images provide unique means of identifying large scale flow-generated lineations produced by former ice sheets. They can be interpreted to reconstruct the major elements which make up the integrated, large scale structure of ice sheets: ice divides; ice streams; interstream ridges; ice shelves; calving bays. The evolving form of the European ice sheet during its decay after the Last Glacial Maximum (LGM) is reconstructed by reference to these components and in the context of a new map showing isochrons of retreat. During the retreat phase in particular the time dependent dynamic evolution of the ice sheet and the pattern of ice stream development are reconstructed.

Crossing lineations are widespread. The older ones are suggested to have formed during molten bed phases of ice sheet growth and preserved by frozen bed conditions during the glacial maximum, particularly in areas which lay, during deglaciation, beneath ice divides and inter-ice stream ridges, both areas of slow flow and frozen bed conditions.

Four phases of growth (A1 to A4) and five phases of decay (R1 to R5) are used to describe the major climatically and dynamically determined stages in the evolution of the ice sheet through the last glacial cycle. The growth and decay patterns are quite different, and associated with major shifts in the ice divide, reflecting growth from the Fennoscandian mountains and decay away from marine influenced margins.

1.0 Introduction

Reconstructions of the form and flow of former ice sheets have largely been based on geological evidence of features such as drumlins, striations and till fabrics which reflect the direction of ice sheet movement, rather than from the more sparse moraines which indicate the locations of former ice margins. The purpose of this study is to improve the glacial geological reconstruction of the European ice sheet presented by Boulton *et al.* (1985) by deducing palaeoflows from a compilation of the distribution, primarily of drumlins, but also of moraines from the area of the ice sheet. The results have been presented in full and interpreted in terms of palaeoglaciological changes by Boulton *et al.* (2001).

2.0 Data

Two principal data sources have been used to plot the distribution of large scale glacigenic landforms within the area of northwest Europe believed to have been occupied by an ice sheet during the last glacial cycle. Satellite images have been used for the inner region of the ice sheet (Fennoscandia, northwestern Russia and much of the area of Estonia, Latvia and Lithuania), and existing fieldwork-based compilations for the peripheral areas (Krasnov, 1971; Chebotareva, 1977; Liedtke, 1981; Kozarski, 1986, 1988; Ehlers, 1990). In the former, MSS images (resolution 79 m) were used, with TM images (resolution 30 m) and SPOT images (resolution 10-20 m) for some more detailed interpretations. Interpretations utilise the capacity of spectral and spatial information to discriminate between drift deposits and bedrock and between glacial landforms and bedrock features (Punkari, 1982, 1985, 1993; Boulton *et al.*, 1985; Boulton & Clark, 1990a, b). The validity of the interpretations of glacial landforms has been checked, in selected areas, against field surveys where they exist but particularly through studies of conventional aerial photography, where there is rarely any doubt whether particular features are drumlins, ice smoothed rock surfaces or moraines. Given the scale of this study however, this can only be done for a very limited number of examples.

Three principle landform types have been mapped:

Eskers. Narrow winding ridges have been identified, approximately parallel to directions of ice flow, which coincide with the locations of eskers marked on the excellent detailed maps of the Norwegian, Swedish and Finnish Geological Surveys (Nordkalott Project, 1986). However, the narrowness and the sinuous or relatively disordered form of eskers makes them difficult to identify on the images used, unless they are long and continuous, in which case their continuity aids their identification. The current mapping therefore underestimates their frequency compared with the results of ground-based survey. Kleman *et al.* (1997) used the distribution of eskers on glaciated surfaces as an index of conditions at the ice/bed interface.

Moraines. Moraines form at and parallel to glacier margins. They are less sinuous than eskers, and can vary markedly in width along their length. This irregularity makes them difficult to identify on our the images. Relatively minor and discontinuous moraines which have been mapped on ground surveys, such as the Middle

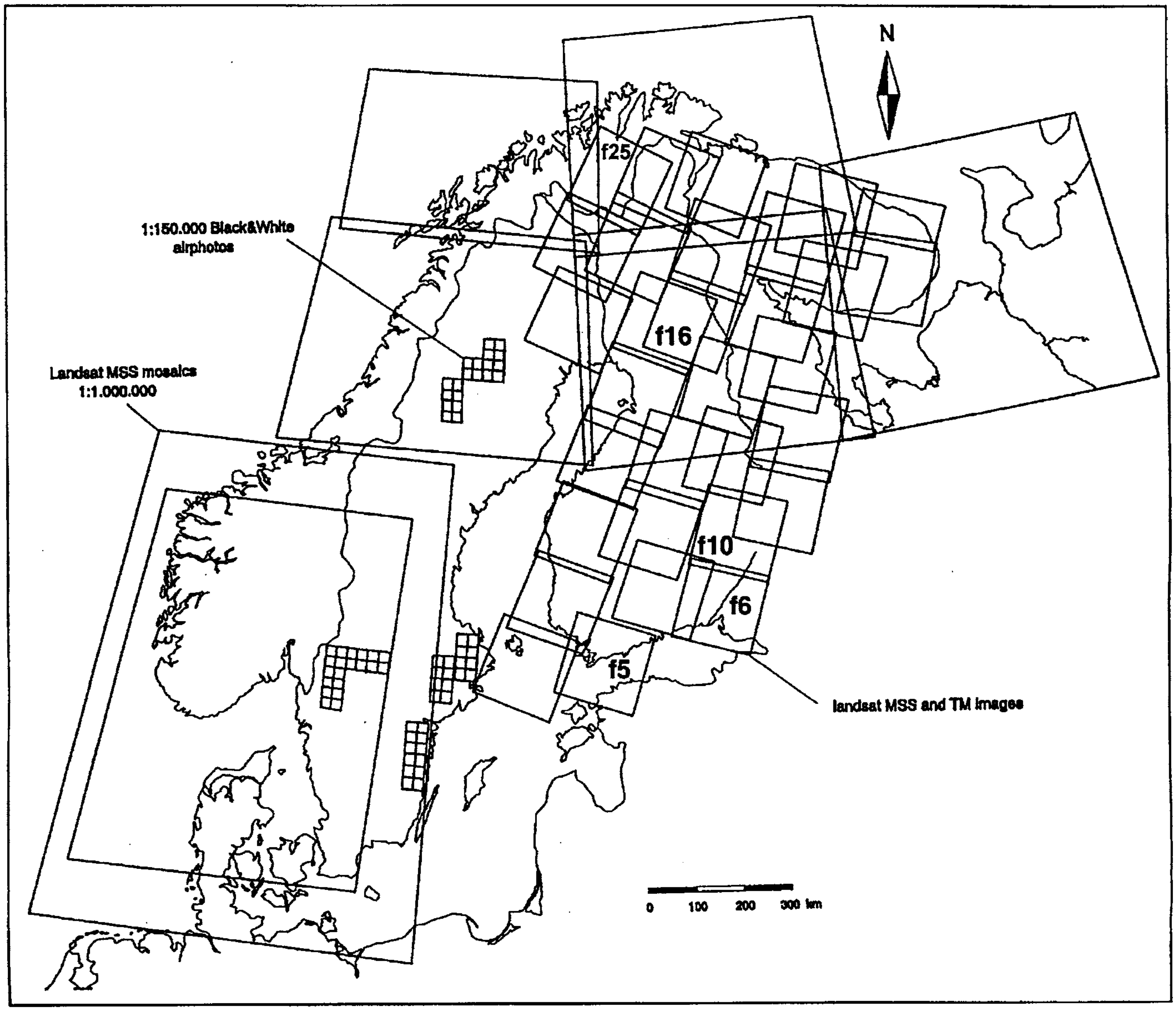

Fig. 1.Location of the satellite images used to map glacial lineations, together with locations of conventional aerial photographs used to check satellite interpretations. Satellite imagery has been used for the whole of the shield areas, where patterns of arable agriculture do not obscure glacial features. Published ground observations have been used to cover areas beyond the shield, and limited areas the Baltic states have been studied using satellite-derived data. Numbers identify images referred to in figures 2 and 3 and in the text.

Swedish Moraines, are difficult to identify, but major or continuous moraines, such as the Salpausselkä Moraines of Finland, are easily recognised.

Drumlins. Satellite survey comes into its own in identifying straight, flow-parallel, streamlined landforms which occur in fields of mutually parallel ridges. They are quite different from the more complex forms of moraines and from eskers. It is thought that almost all such flow-parallel features that we have mapped are sediment ridges, 10s-1000s of metres in length, metres to 10s of metres in height, and 10s to 100s of metres in width. They show great diversity of size and of detailed form and can all be reasonably grouped together as 'drumlins'. Satellite image-based maps tend to show a greater density of drumlins than maps of drumlin fields compiled from field surveys. Although such comparisons show that small drumlins are missed by satellite surveys, we are often able to identify drumlins of great extent and low relief which are too large to be discerned by field surveys, suggesting that in many drumlin fields, satellite surveys produce a better estimate of the total drumlin population. Satellite survey is also an effective means of identifying where changing directions of ice flow have created two populations of drumlins, in which older drumlins have been partially reworked to create a second, cross-cutting drumlin set (Boulton & Clark, 1990b).

The locations of the images, on which the present interpretations are based, are shown in Fig. 1. Examples of images and interpretations made from them are shown in Figs 2-4. Satellite image survey has the advantages that a

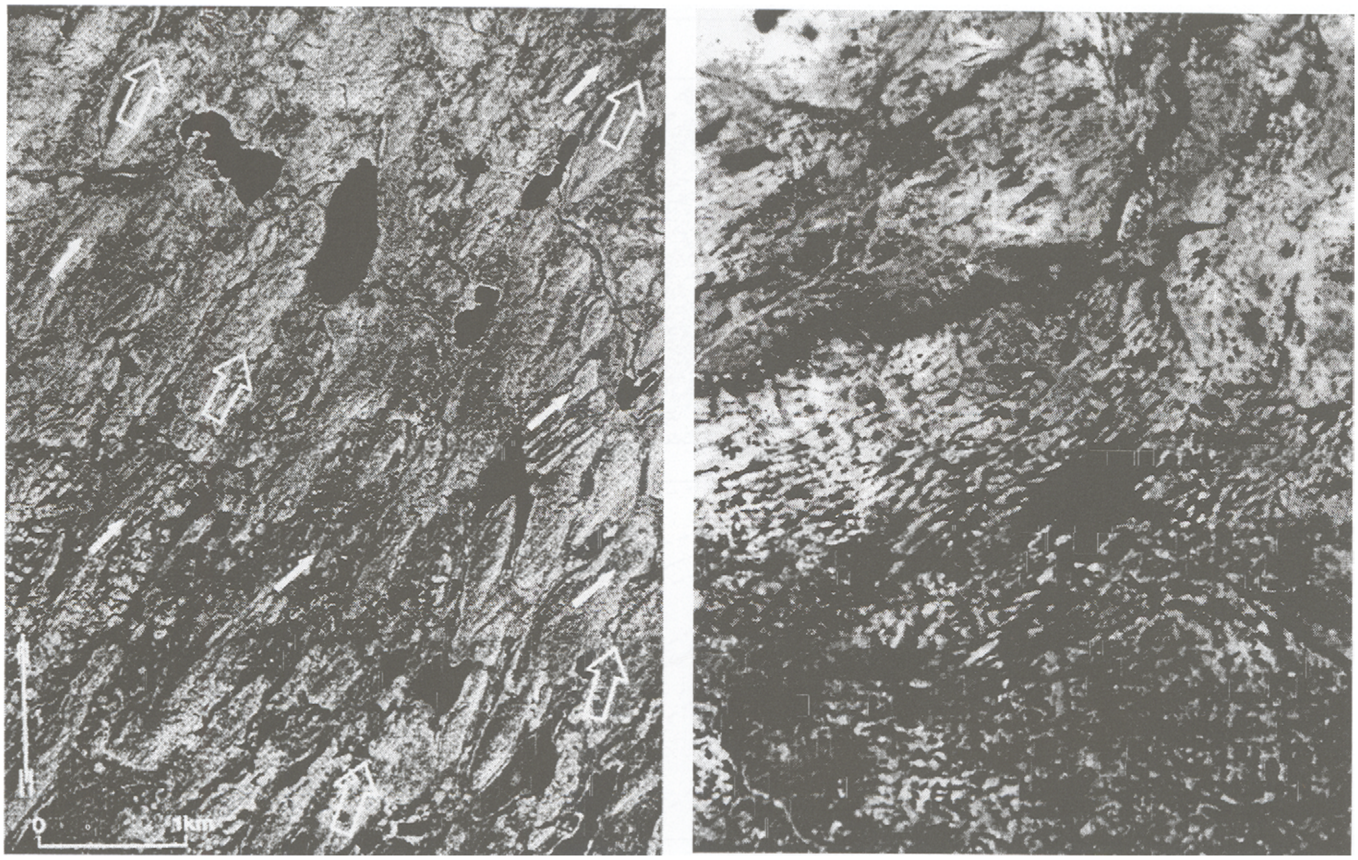

Fig. 2a-b. a) Conventional aerial photograph showing crossing lineations at Utsjoki in northern Finland. Drumlins directed towards the NNE (trend shown by large open arrows) have smaller flutes and drumlins directed towards the NE superimposed on their surfaces (trend shown by smaller filled arrows), suggesting a clockwise rotation of ice movement of about 25°.

b) A satellite image immediately to the south of a), from f25 on Fig. 1. The upward direction is towards 310° and the width of the image is 130 km. A strong S-N lineation defined by an extensive drumlin field is shown in the central part of the image, where it overprints an older NNE-trending lineation.

single operator or a small group can apply a well-defined set of criteria on a continental scale; that large features that are often unresolveable by ground or air photo survey can be identified and that groups of individually indistinct linear elements can be recomposed visually into coherent patterns to infer integrated patterns of ice flow. The level of coherent detail about patterns of lineation over wide areas that can be mapped from satellite imagery is generally greater than has hitherto been obtained from ground surveys such as those used by Kleman *et al.* (1997; see their Fig. 3). Satellite-derived data contrasts with stratigraphi-cal and sedimentological evidence of palaeoglaciology, which tends only to yield information from very small areas about continental-scale glacier behaviour. In this study continent-wide spatial patterns of flow parallel lineations have been plotted as a basis for reconstructing spatial patterns of ice sheet behaviour. It has the disadvantage that it only permits, at best, relative dating and has the danger that temporally separated flow events are grouped together. It should ultimately be integrated with stratigraphic and sedimentological data (Dongelmans, 1997).

To guard against subjective interpretations, the authors have individually and independently produced interpretations of the same areas and cross-checked the results. In general, substantively different interpretations have not been found. In some areas, bedrock structures show linear patterns, but many such areas can be eliminated by using existing geological and geophysical maps.

Figs 3a, c and e show details of flow-parallel lineations observed on satellite images at a scale of 1: 300,000. We term these first order lineations. In some cases it is clear that different parts of an the image show flow lineations which could not reflect synchronous patterns of basal ice flow. In some cases, lineations cross, and in some of these, it is possible to establish the relative ages of lineations from satellite images (Boulton, 1987). In some, relative ages have been checked by reference to aerial photographs, but in others no clear interpretations of relative age could be made. It is not possible to show first order interpretations on an ice sheet wide basis in figures of the scale used in this article. In order to show ice sheet wide patterns, three successive approximations have been made:

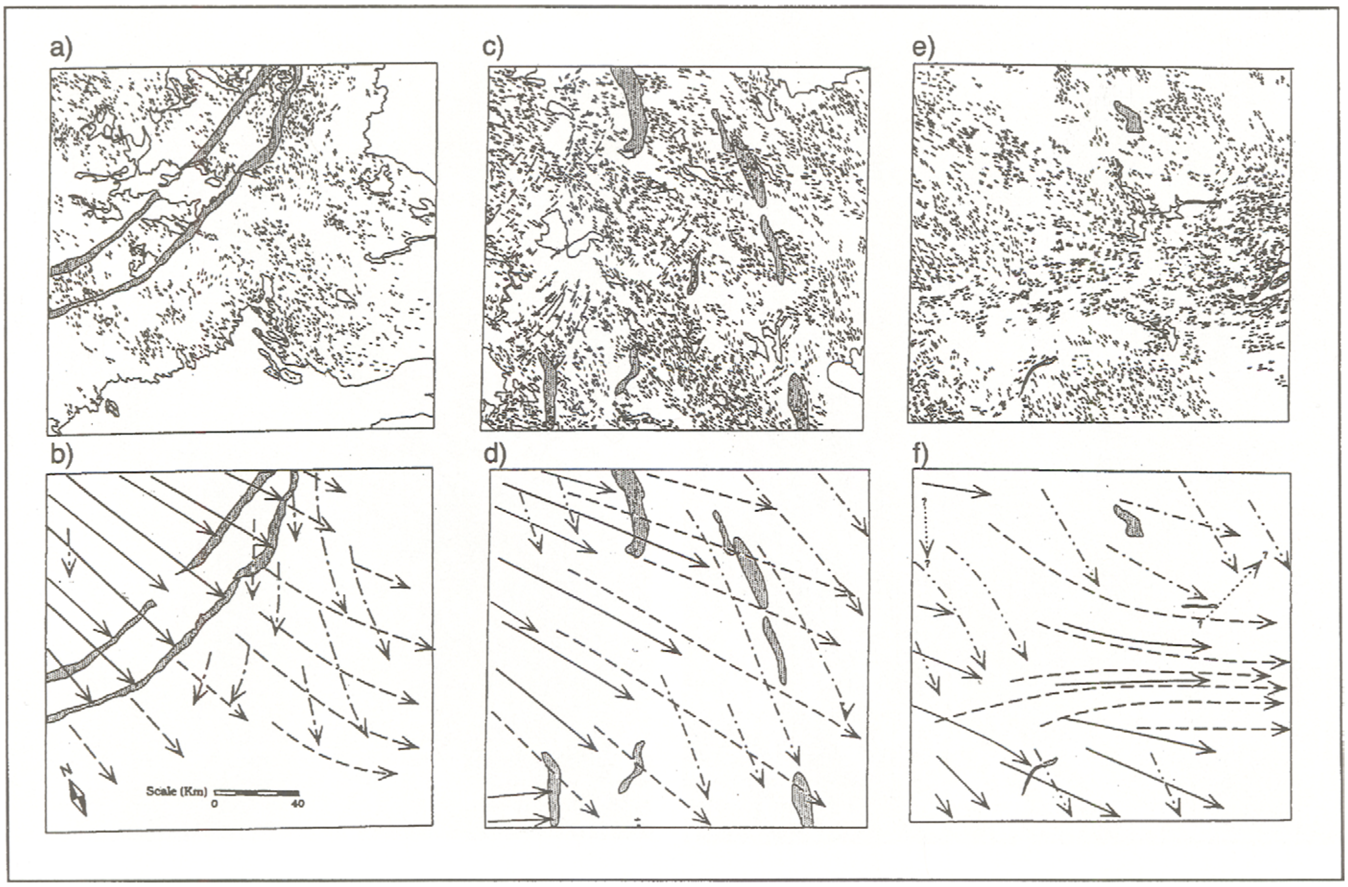

Fig. 3 a-f: Examples of interpretations of satellite images. a), c), and e) show first-order interpretations directly from satellite images. b), d) and f) show interpretations of the pattern in terms of integrated, second-order lineation sets. Different second order sets are shown by different arrow types. ***a-b)*** *Image f5 (Fig. 1) in the area of the First and Second Salpausselkä Moraines in eastern Finland. The interpretation suggests an oldest fanning lineation pattern (possibly an ice stream) reflecting a NNE-SSW flow, overlain by successively younger lineations reflecting NNW-SSE flow. The NNE-SSW sets south of the First Salpausselkä Moraine are separated because of the long ice sheet still-stand at the moraine.* ***c-d)*** *Image f10b (Fig. 1) in eastern Finland. An older NS lineation with a superimposed NW-SE lineation. Earlier (dashed line) and younger integrated sets are distinguished.* ***e-f)*** *Image f16 (Fig. 1) from northern Finland showing an early N-S lineation superimposed by an older and a younger WNW-ESE lineation, convergent towards the SE, reflecting a change from non-streaming flow to streaming flow towards the head of the White Sea during ice-sheet retreat.*

a) Maps of first order interpretations have been generalised to produce second-order interpretations of lineation sets which clearly form parts of spatially integrated groups and can be summarised by a series of lines which are longer than the original lines representing individual geomorphological features (Fig. 3b, d, f). Many show several crossing lineation sets.

b) The second order interpretations have been further compiled to produce third-order lineation maps of large sectors of the ice sheet (Fig. 4).

c) These have been further simplified to produce fourth-order maps of lineation trends over the whole area of the ice sheet (Fig. 5) by selecting representative third order lineations trends.

3.0 The Last Deglaciation

3.1 Geological reconstruction - the pattern of retreat

Two types of data are used in reconstructing the pattern of retreat. From the area of the Shield, flow lineations and a relatively small number of ice-marginal moraines observed on satellite imagery are used. In the flanking area of younger sediments, a compilation of field observations from a variety of authors and a relatively small cover of satellite images from the Baltic states and north-west Russia and Belorussia are utilised.

Figs 3-5 show that from the area in which satellite images have been used, flow lineations from different phases of ice sheet history are preserved and superimposed. In this section, those lineations are identified which are thought to have been produced during the last deglaciation. Section 5 shows how they can be stripped away in order to infer ice sheet flow patterns from earlier stages. Ice-marginal landforms (end moraines, outwash plains, etc.), produced during deglaciation, are direct evidence of the pattern of final retreat (Fig. 6). However, except for the Younger Dryas moraines, only a few significant end moraines occur in the shield areas to guide reconstruction of deglaciation. Following rules set out in section 3, it is expected

- that the strongest lineation to be generated in the sub-marginal zone of the ice sheet where velocities were highest,

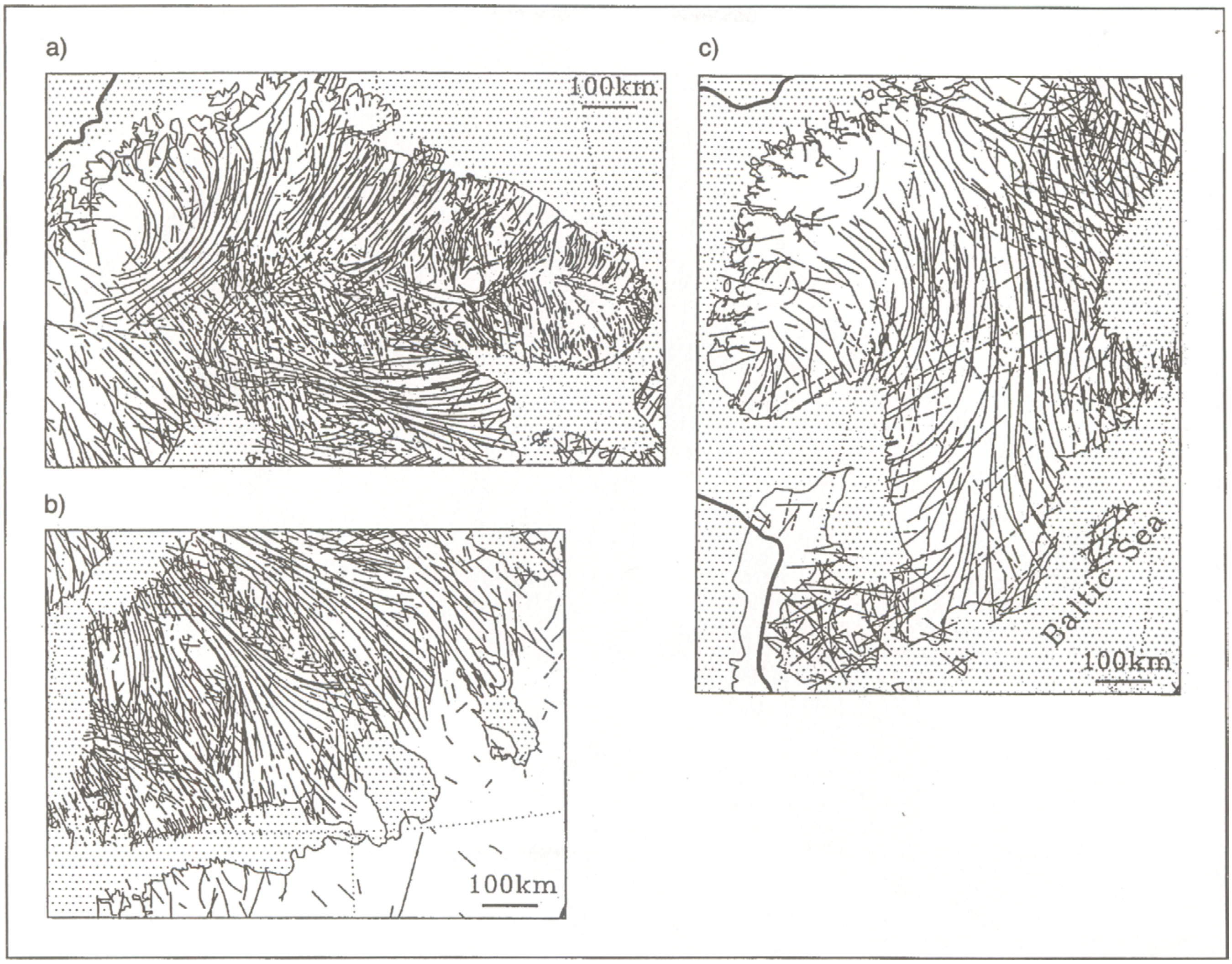

Fig. 4a-c. Third-order interpretations, generalised from second order lineations such as those shown in Fig. 3, showing lineation patterns over the shield area of the ice sheet. a) Northern Scandinavia. The lineation convergences into the head of a very strong lineation sets running ESE towards the White Sea, which shows divergence along the White Sea shore are particularly interesting. b) Southern Finland and the area around the Gulf of Finland. Note the crossing lineations in the inter-ice stream areas and reflecting interactions between ice streams during their lifetimes. c) Southern Sweden, southern Norway and part of Denmark.

- that in most cases, the greatest lineation density will reflect flow in this zone during the retreat phase,
- and that, on a horizontal bed, the trend of these lineations will lie normal to the trend of the ice margin.

The prime exceptions to this will be:

- where fast ice streams have produced strong lineations far from the ice sheet margin and which are preserved because streaming flow ceases prior to deglaciation of the area.
- where lineations are created in a zone of melting and which are preserved because of onset of ice/bed freezing prior to deglaciation.

If these rules are applied to lineations in areas where ice-marginal moraines have survived, it is found that the ice-marginal retreat pattern inferred from the densest lineations coincides with that inferred from moraines (eg. Fig. 3a-b). This approach permits:

- a definition of the general pattern of retreat with much greater precision than by using ice margin-parallel forms alone;
- an identification of those lineations which are not normal to the retreating ice margin, and which thus reflect either flow patterns far from the retreating ice margin or patterns produced during earlier phases of glaciation.

In Fig. 7, following Boulton *et al.* (1985), ice margin trends during retreat have been determined by assuming that they lie normal to the dominant flow lineations. This gives a clear picture of the trajectory of retreat of the ice margin and of the sub-marginal flow patterns associated

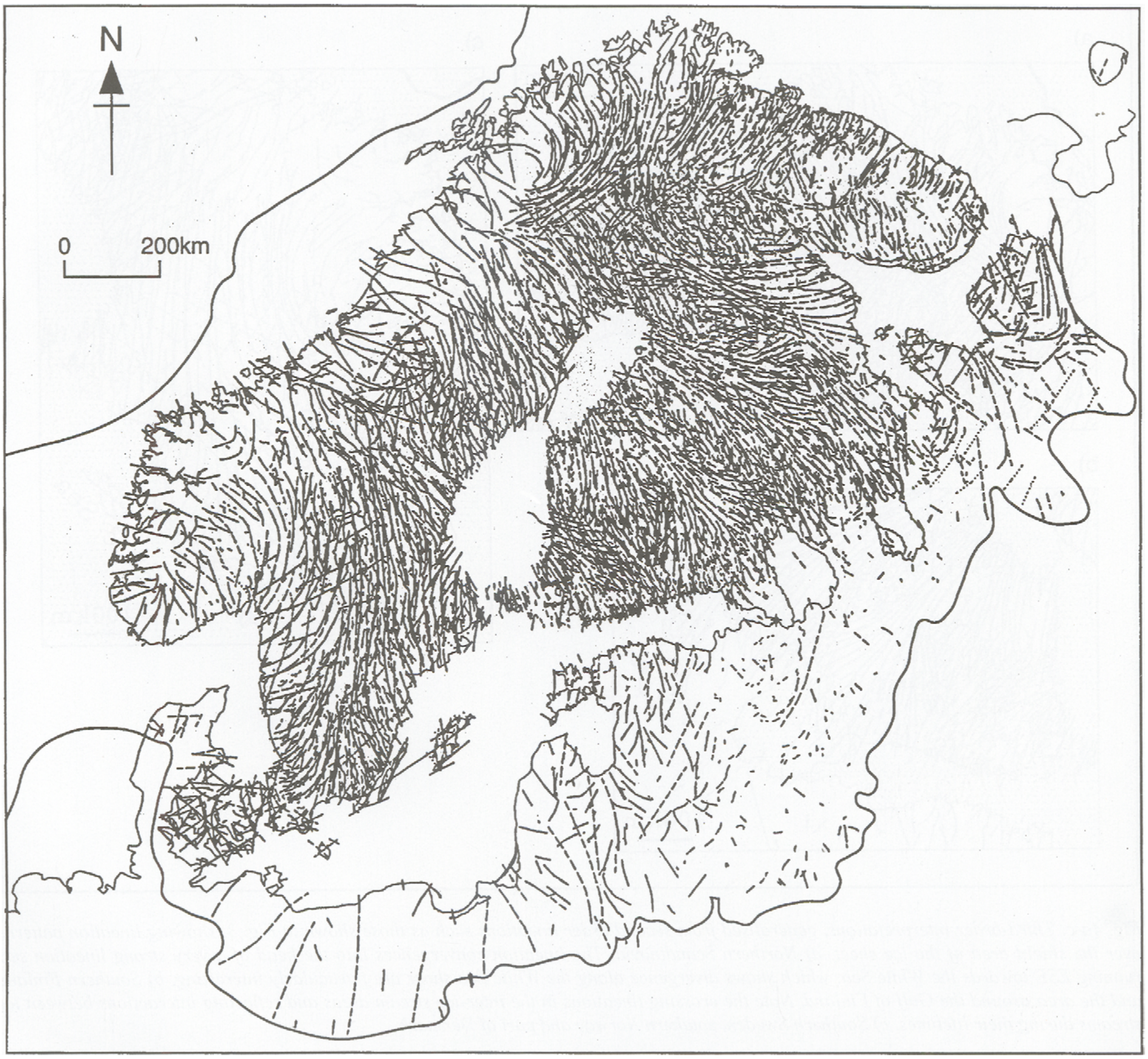

Fig. 5. Fourth-order longitudinal lineation sets plotted on a European scale. Note the areas of abundant crossing lineations between the strongly lineated zones which are interpreted as ice streams, particularly clearly seen in southern Finland (also Fig. 4b). They reflect the location of sluggish inter-stream ridges. They may also be characterised by basal freezing. Crossing lineations also occur in former ice divide areas on the eastern side of the Scandinavian mountain chain.

with it. It permits interpretation of aspects of the dynamic behaviour of the ice sheet during retreat.

Beyond the Shield, in the area south of the Baltic Sea, much of the evidence of ice sheet behaviour during retreat comes from the distribution of major morainic ridges formed at the maximum and during retreat. The maximum extent of the ice sheet and its retreat in the northeastern North Sea and on the Norwegian continental shelf have been summarised by Holtedahl (1993) and Sejrup *et al.* (1994).

3.2 Geological reconstruction – the tempo of retreat

In this section an attempt is made to translate the pattern of retreat shown in Fig. 7 into one showing specific iso-chrons (Fig. 8). This is relatively simple in the shield area where a unique record of the tempo of ice-sheet retreat is contained in the annually laminated varve chronology of Sweden, Finland and the area around the Gulf of Finland.

Beyond this zone however, lineation data is sparse and chronology is much more difficult to establish. This latter

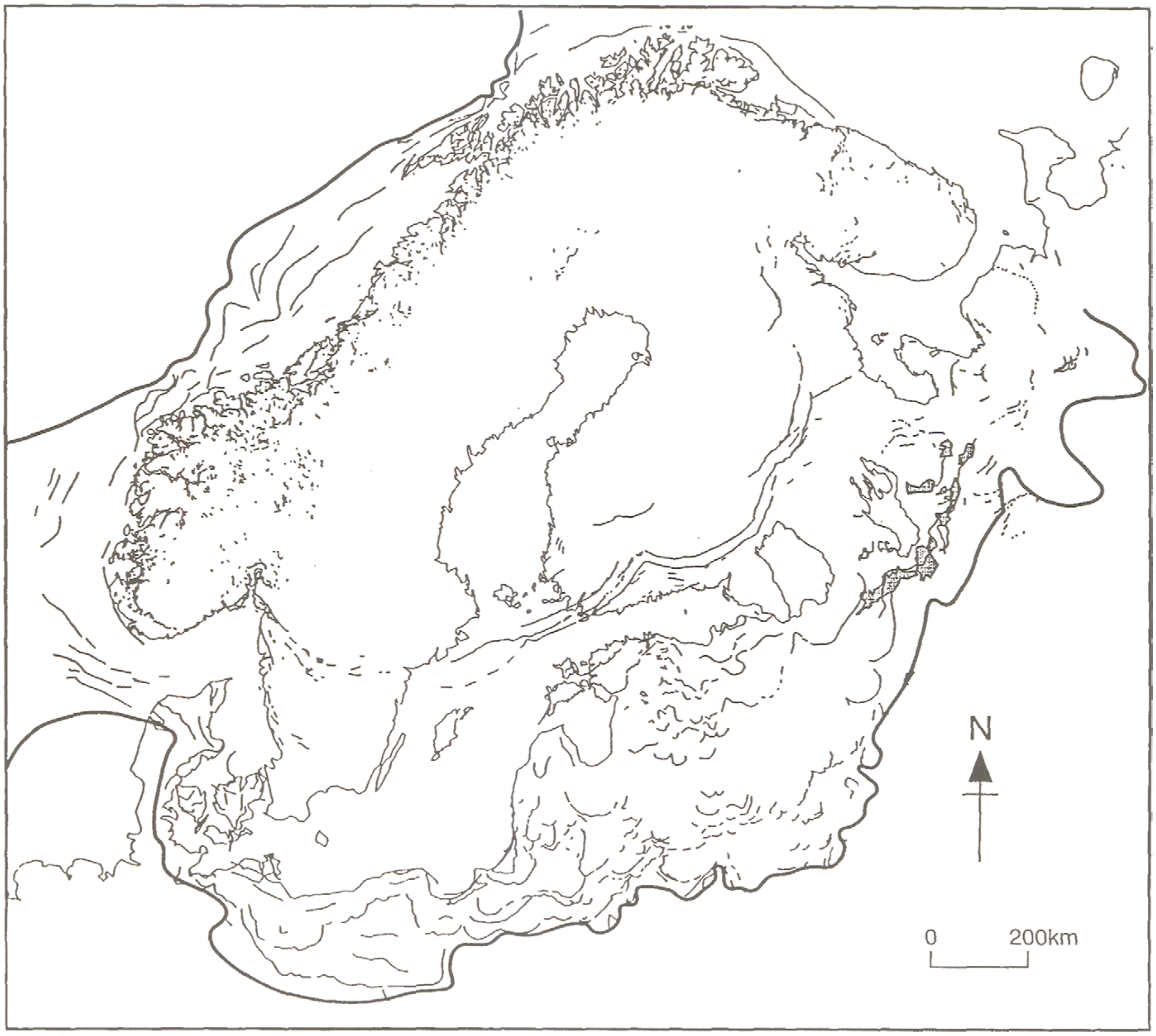

Fig. 6. The principal zones of ice-marginal moraines and other ice-marginal deposits within the area of the Weichselian ice sheet. Apart from the Younger Dryas Salpausselkä moraines and their correlatives, most moraines occur within the zone of relatively slow retreat lying beyond the shield (cf. Fig. 7).

area is discussed before retreat chronology within the varve zone is analysed.

3.2.1 'Pre-varve' retreat

There is strong evidence that the maximum extent of the Late Weichselian Scandinavian ice sheet was time-transgressive. The earliest advance to the maximum for which evidence is available is in the southwest, where the ice sheet, flowing from a centre in southwest Norway, crossed the North Sea to become confluent with an ice sheet over the British Isles at about 28 ka (Sejrup *et al.*, 1994, 2000) (Fig. 8). There is evidence that during this phase, the Norwegian Channel, which extends from the Skagerrak and skirts the southern and southwestern coast of Norway as far as the continental shelf edge off southwest Norway, was the location of a major ice stream (King *et al.*, 1996; Sejrup *et al.*, 1996, 1997).

The connection between the European and British ice sheets was broken shortly after 23 ka, and the European ice sheet margin retreated to the vicinity of the coast of southwest Norway by about 20ka (Valen *et al.*, 1996). Between 20 ka and 18 ka, the ice sheet readvanced from the vicinity of the Norwegian coast to reach a maximum just beyond the southwest margin of the Norwegian Channel (the Tampen advance; Sejrup *et al.*, 1994). It seems most likely that the sharp re-entrant in the maximum extent of

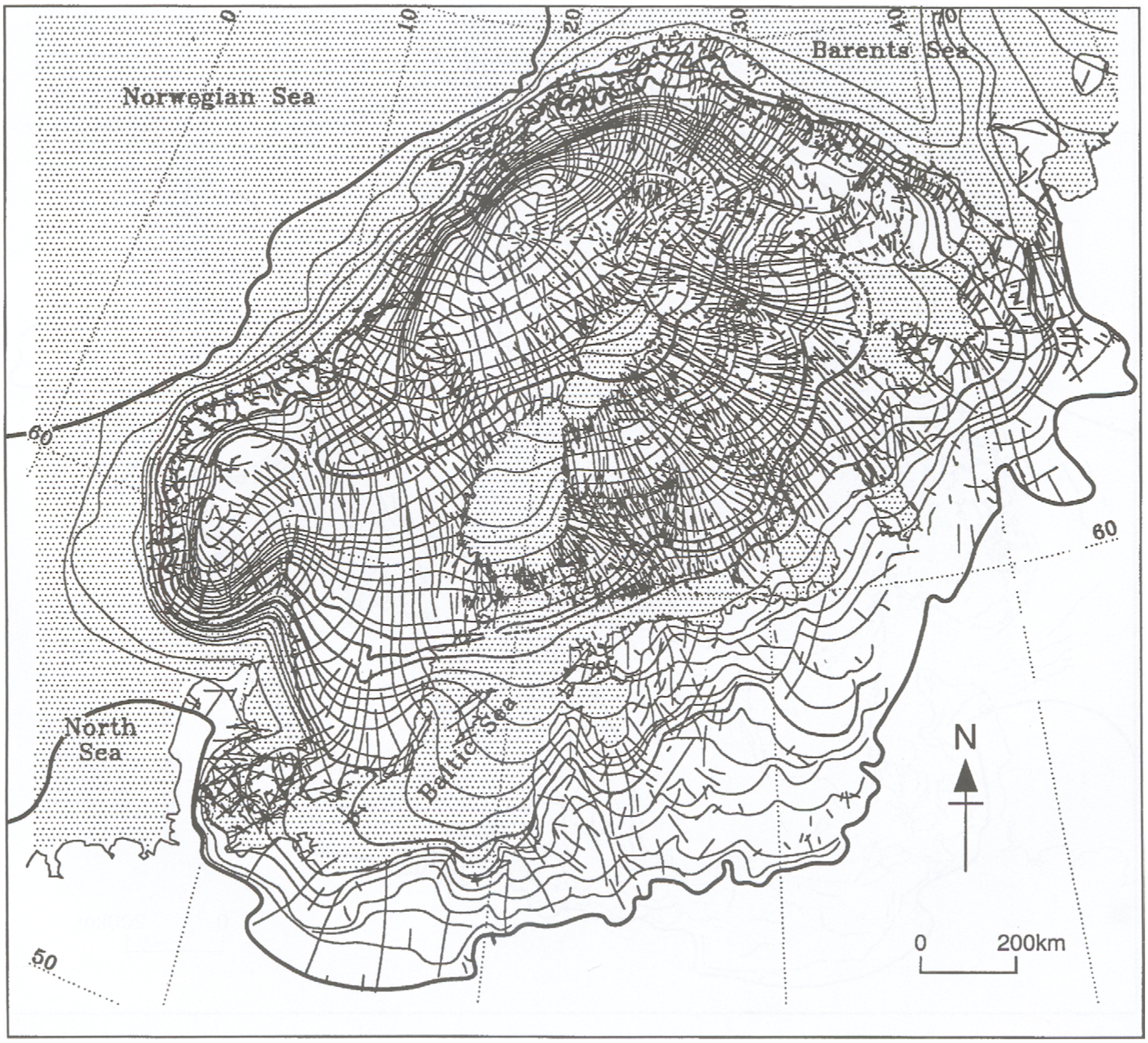

Fig. 7. Inferred pattern of ice-front retreat of the last Northern European ice sheet, superimposed on the dominant patterns of longitudinal lineation. Within the area of the shield, retreat lines are constructed from normals to fourth-order, flow-generated lineations (Fig. 5). Beyond the shield they are largely derived from the distribution of moraines (Fig. 6). Within this latter area, some longitudinal lineations are also shown, derived from published field mapping of drumlins. Note that there is a strong correspondence between inferred ice margin positions on the shield and the few moraines, such as the Salpausselkä, which occur there.

the Weichselian ice sheet margin in northern Jutland (Fig. 8) represents an overstepping by this advance of the margin of the older North Sea ice sheet, such that the maximum extent in much of Jutland and northwest Germany, is roughly contemporary with the Tampen advance. Furthermore, the distribution of moraines in northern Germany suggests that the eastward continuation of this advance, represented by the north-south Weichselian ice margin through Denmark, also oversteps the line of the older, Brandenburg moraines and continues into the Poznan Moraines (Fig. 8).

The configuration of moraines in northern Germany and Poland (Fig. 8) suggests that an early advance occurred in this sector to the glacial maximum along the Brandenburg-Lezno moraines, and that this was followed by a retreat and then a readvance which created the Poznan stage moraines (Kozarski, 1986, 1988).

Further to the east and west, the readvance formed the glacial maximum. It is possible that the early, Brandenburg-Lezno advance was a consequence of stronger ice sheet flow down the axis of the southern Baltic Sea. It may reflect the same phase of ice sheet growth which led to

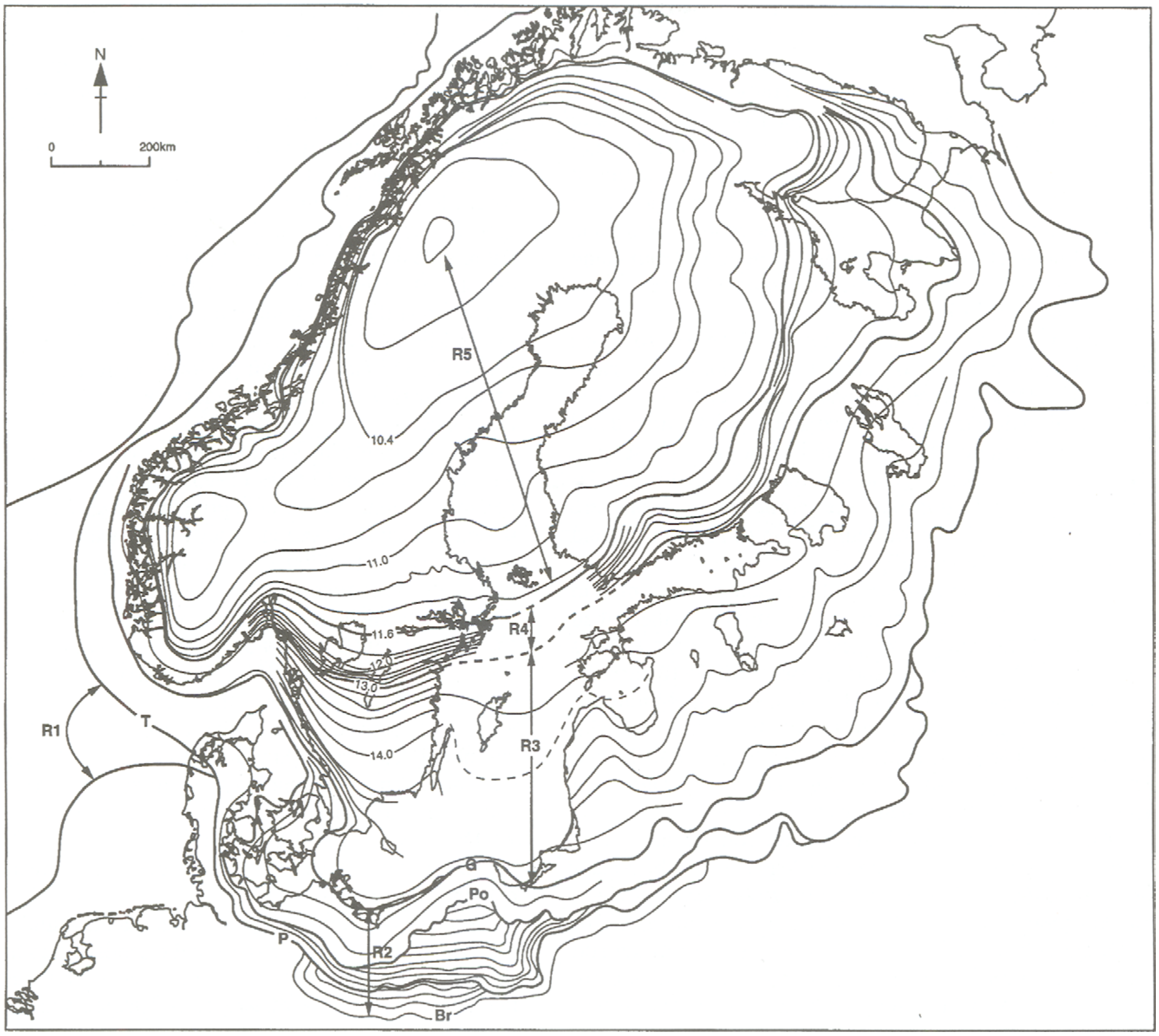

Fig. 8. Isochrons of retreat of the last Northern European ice sheet. The retreat pattern shown in Fig. 7 is translated into isochrons of retreat by calibration with the Swedish and Finnish varve timescales and available corrected ^{14}C *determinations. Retreat is subdivided into a series of principal phases R1-R5, with dates of R1 = 28 ka-20 ka; R2 = 20 ka to 15.2 ka; R3 = 15.2 to 13 ka; R4 = 13 ka to 11.5/6 ka; R5 = 11.5/6 to final decay. Dating of the early stages of retreat is speculative. The major LGM and early retreat moraine phases are shown as Br = Brandenburg-Lezno (22-20 ka); P = Poznan (18.0 ka); Pm = Pomeranian (16.5 ka); G = Gardno (15.4 ka). The pattern of retreat in Poland follows Kozarski (1986). It is clear that the maximum glacial extent was reached earliest in the southeast (at about 28 ka – Sejrup et al., 1994) and last in the east (between 17 and 15 ka). The ice sheet was still advancing in the east whilst retreating in the west.*

extension of ice from the southern Norwegian centre of ice sheet growth into the North Sea.)

The later, Pomeranian Moraines (Fig. 8), have been shown by Kozarski (1986) to cut across the earlier, post-Poznan Moraines in Poland, representing yet another major advance of the ice sheet margin in this sector. The Gardno Moraine in northernmost Poland (Kozarski, 1986) may represent a relatively small readvance, as there is no clear evidence of major cross-cutting relationships with the major moraines to the south.

There is very little independent evidence of the age of the glacial maximum or of ice-marginal positions in the southern (German-Polish) sector of the ice sheet. Prior to its maximum extent, the expanding ice sheet moved across the southern Norwegian and southwest Swedish coasts between 30 and 24 ka (Hillefors, 1974; Andersen, 1987) and crossed the Polish coast slightly before 22ka BP (Kozarski, 1988). The ice sheet is suggested to have advanced to the maximum extent in Poland shortly after 20 ka BP (TL yr - Ralska-Jasiewiczowa & Rzetkowska, 1987; Mojski, 1992). The time window for the glacial maximum in Poland is closed by an uncalibrated ^{14}C date of 13,500 BP on the northern coast of Poland on the earliest organic deposits after deglaciation (Ralska-Jasiewiczowa & Rzetkowska, 1987). Adjusting this date using the correction of Bard *et al.* (1993), gives an age of 15,950 BP (from here on ^{14}C ages are given as corrected ages unless otherwise stated). Although there is considerable uncertainty about the timing of the glacial maximum and of deglaciation of the area to the south of the Baltic Sea, the above estimates are not inconsistent with other records. The isotopic minimum in deep-ocean stratigraphy, thought to record the global ice volume maximum during the last glacial period, occurred at 22 ka (Imbrie et al., 1984), as did the period of maximum cooling in the Greenland ice-core record (Johnsen et al., 1992; Bond et al., 1993).

Support for the general distribution of retreat isochrons in the southeast sector of the ice sheet during the early stages of retreat is given by Sandgren *et al.* (1997), who have used palaeomagnetic correlations with varve-dated sites in southern Sweden and Karelia to suggest deglaciation of Lake Tamula in Estonia at 14,400 calendar years BP. It indicates that the proposed isochrons are consistent with the time of deglaciation of northern Poland and of southern Estonia (Fig. 8).

If the ice sheet readvances in the southern sector margin (Kozarski, 1988) are climatically driven rather than dynamic, it is tempting to suggest the following correlations with post-LGM cooling events in the GISP/GRIP records: Poznan Moraine - 18,500 BP; Pomeranian Moraine - 16,500 BP; Gardno Moraine- 15,400 BP; compared with Kozarski's (1986) estimates of 20,400 for the Lezno, 18,400 for the Poznan, 15,200 for the Pommeranian and 13,300 BP for the Gardno moraines.

The trend of morainic features in the southeastern sector of the ice sheet suggests that the features which are co-linear with the Gardno Moraine ultimately converge, about 300 km south of Lake Ladoga, with the maximum extent of the Weichselian glaciation (Fig. 8), just as the Poznan and Pommeranian moraines appear to further west. If this is so, it suggests that whilst the southern and south western margins of the ice sheet were retreating by up to 300 km, the eastern and southeastern margins of the ice sheet were stationary or still advancing. This conclusion is supported by evidence that in the area to the southeast of the White Sea, the glacial maximum was not reached until after 17 ka and that retreat from the maximum position did not begin until about 15 ka (Larsen *et al.*, 1999).

In the northern sector, the ice sheet appears to have reached a maximum in northern Norway between 19-18.5 ka BP (Vorren et al., 1988), whilst an ice sheet over the Barents Sea, believed to have been confluent with the European ice sheet, is thought to have reached its maximum extent and begun to collapse rapidly by 15 ka and had largely disappeared by 12 ka (Landvik *et al.*, 1998).

3.2.2 The varve zone

The most important record of the tempo of deglaciation is in the varve-based 'Swedish Timescale' (De Geer, 1940), which has been extended back continously from final deglaciation at about 8,700 varve years B.P. to about 11,500 varve years BP (Strömberg, 1985). However, Younger Dryas-age advances of the ice sheet in the Middle Swedish End Moraine zone make it very difficult to connect this chronology with that represented by about 2000 varve years in southern Sweden. The latter is a floating chronology because of dislocation in the Middle Swedish End Moraine zone, but if this dislocation is ignored, appears to extend back to about 13,300 varve years BP (e.g. Wohlfarth *et al.*, 1995; Strömberg, 1994; Björck *et al.*, 1995). The work of Strömberg (1994) appears to indicate that the maximum of the Younger Dryas readvance in the Middle Swedish zone was reached at 11,410 varve years BP, and that ice front positions representing a period of 180 years were later overridden by the Younger Dryas readvance.

The interpretation summarised by Björck *et al.* (1995) and Wohlfarth and others (1995; and Wohlfarth personal communication, 1995) can in principle be used to determine the rate of ice-sheet retreat from 13,258 varve years BP, in eastern Scania, to 9200 varve years BP in Västernorrland. Although correlation between varve and ^{14}C years has been difficult to achieve, a direct comparison of varve and AMS ^{14}C chronology (Björck *et al.*, 1995) shows the latter to be systematically older by 500-1,500 years during the period 9,500-12,500 varve years BP, although there is good agreement between the varve chronology and the ^{14}C chronology calibrated (Stuiver & Reimer, 1993) against the German oak-pine chronology (Kromer & Becker, 1993) and U/Th ages on corals (Bard et al., 1993) up to about 12,000 varve years BP. The U/Th-calibrated ^{14}C record beyond 12,000 varve years BP gives much greater ages than the varve record.

A means of confirming the precise magnitude of the mis-match between varve years and calendar years has been developed by Andren *et al* (1999). They have suggested that varve thickness will largely reflect ice sheet melting rate as a direct result of changing summer temperatures. They have compared varve thickness data for the period 10,250-11,550 varve years with the GRIP ice core ^{18}O record (Johnsen *et al.*, 1992), which is probably the best proxy, high resolution guide to the trend of climatic variations in western Europe during the period, reflected for example in the excellent match between the GRIP record (Fig. 9a) and the sea surface temperatures in the North Atlantic (Fig. 9b; Kroon *et al.*, 1997). They show that a very good correlation exists between the two records for the highly distinctive pattern of change during the Younger Dryas/Preboreal climatic shift, provided that 875 years are added to the varve ages.

We have therefore added 875 years to the whole varve chronology, and added 180 years to that part lying south of the Middle Swedish End Moraines, to produce the pattern of ice-sheet retreat rate shown in Fig. 9c. The pattern is very similar to that shown by the GRIP and NE Atlantic records, but lags behind it. The best fit is obtained by assuming that the retreat rate lags 250 to 500 years behind phases of climatic warming.

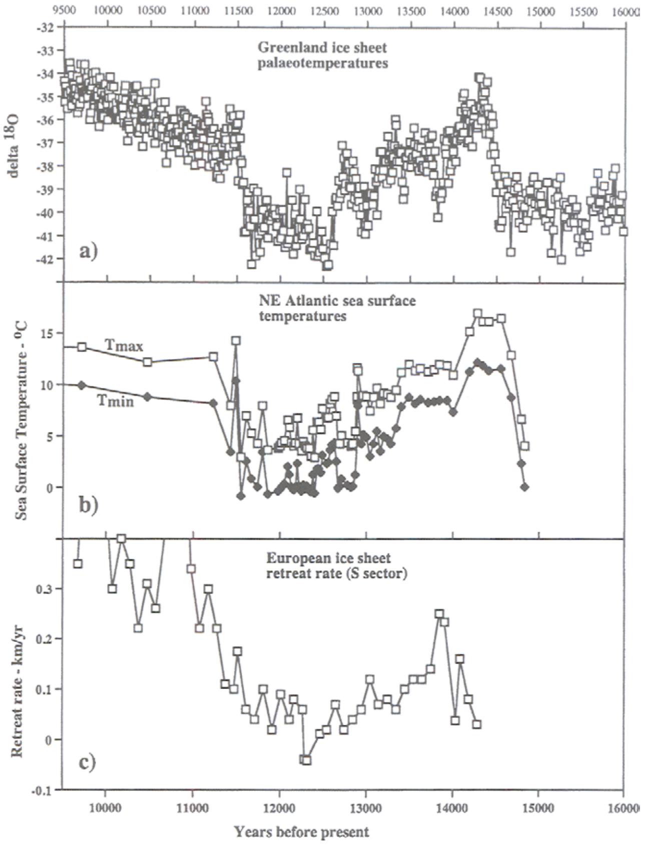

Fig. 9 a-c. Correlation of the corrected varve zone retreat of the ice sheet with other contemporary records of climate change. a) The GRIP $\delta^{18}O$ record (Johnsen et al., 1992) between 16,000 and 9,500 BP. b) Sea surface temperture record in the NE Atlantic, off western Scotland (Kroon et al., 1997) showing the match between a European proxy climate indicator and the GRIP record. The upper curve shows maximum and lower curve minimum temperatures. c) Retreat rate of the European ice sheet in the Swedish varve zone. The timescale is derived from a correlation with the GRIP core based on variations of varve thickness (Andren et al., 1999). It suggests that the retreat rate lags climate forcing by 250-500 years. This is suggested to be an ice-sheet dynamic effect.

This corrected varve-based chronology is now applied to the ice-sheet retreat pattern shown in Fig. 8 by ascribing ages to retreat isochrons at the points at which they cross the varve transect series in eastern Sweden. Fig. 10a shows the inferred pattern of time dependent retreat of the southern sector of the ice sheet, not only for the area to which the corrected varve timescale can be applied, but also for the earlier phase of retreat by using the tentative correlations offered between the Poznan, Pommeranian and Gardno readvances and Kozarski's (1986, 1988) estimate of retreat rate between them.

A varve chronology for Finland was first published by Sauramo (1918), but has been a 'floating chronology' until recently. Strömberg (1990) has suggested that the zero varve of the Finnish chronology (Sauramo, 1923) was equivalent to 10,643 varve years in the Swedish chronology. This has been used as the basis of the chronology in Finland. It has now been extended to the area of the eastern Gulf of Finland and as far as Lake Ladoga (Hang, 1997).

Although it is still premature to create an unequivocal reconstruction of European ice sheet deglaciation, in view of much poor or heterogeneous dating evidence, the absence of evidence of ice margin trends in marine areas, and the fragmentary evidence of trends in areas such as Belorussia, it is helpful to create a deglaciation hypothesis

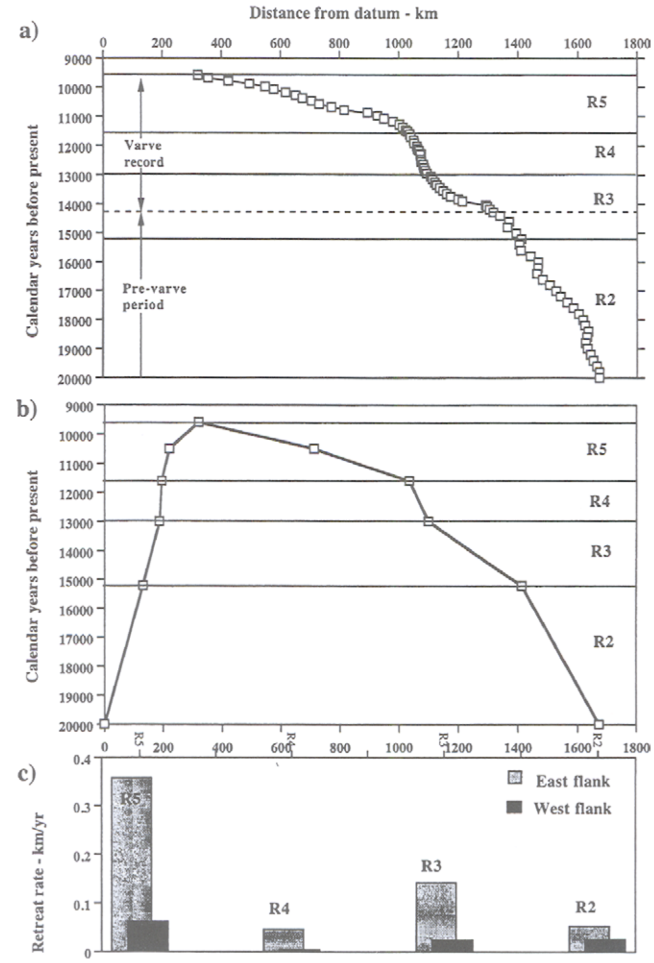

Fig. 10 a-c. Tempo of retreat of the ice sheet along a transect from northern Germany, along the southeast coast of Sweden and west to the Norwegian continental shelf (Boulton et al., 2001). a) Retreat along the transect on the southern flank of the ice sheet. The early part of the retreat is based on estimates of age by Kozarski (1986, 1988) and speculative correlations with the GRIP record. After 15.2 ka, the retreat chronology is based on the varve chronology as far as the north of Stockholm. Beyond this point, varve dates are projected onto the line of section using the isochron map (Fig. 8). b) Net retreat during each of phases R2-R5 along the transect on the eastern (right) and western (left) sides of the final ice divide. The retreat on the western flank of the ice sheet prior to 13 ka (Younger Dryas Stadial) is poorly known. c) Retreat rates during phases R2-R5 on the eastern (light shading) and western (dark shading) flanks of the ice sheet. Retreat rates are invariably less on the western side of the divide, reflecting less negative mass balances on the more maritime and mountainous flank.

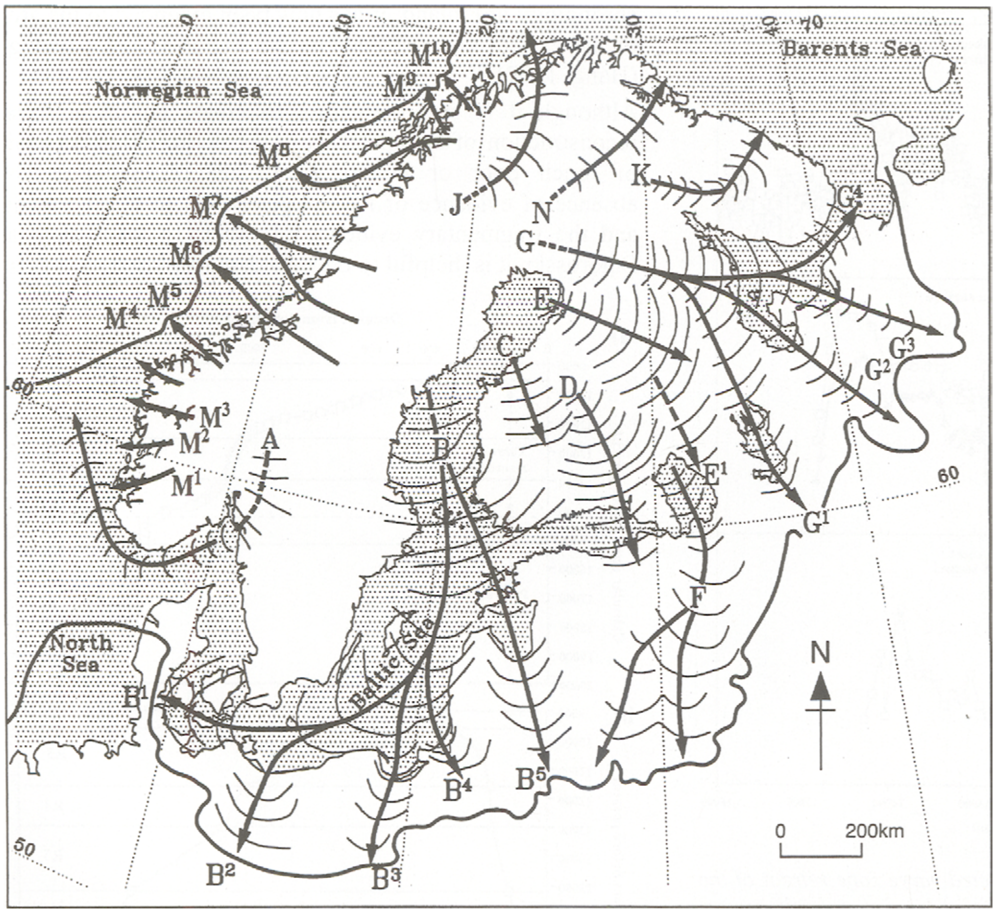

Fig. 11. Time-dependent trajectories of land-based major ice streams which existed during decay of the ice sheet, marked A-J (named in the text). The stippled zones show the swathes along which ice streams were active during deglaciation. The swathes are presumed to reflect time-transgressive activity, not the length of an ice stream at any one time. Marine ice streams are shown on the western margin of the ice sheet. Evidence for these elements include shelf troughs, arcuate moraines at the mouths of shelf troughs and major accumulations of glacial sediment at the shelf edge or on the upper continental slope.

as a spur to testing and to provide a best estimate of deglaciation for quantitive simulation models. The evidence of trends of ice margins and the varve chronology in much of Sweden, and to a lesser extent in Finland, make it possible to produce a persuasive reconstruction of deglaciation in those areas. South of the Baltic and the Gulf of Finland and in northern Finland and the Kola Peninsula, the dating evidence is sparse, and ice-marginal trends are uncertain east and southeast of the Gulf of Finland. However, combining the concentric trends normal to flow lines shown in Fig. 7 with the evidence of ice-marginal landforms and the chronological information discussed above, allows the reconstruction of the tempo and pattern of ice sheet decay during the last deglaciation in Fig. 8. The interpreted tempo of deglaciation along a transect from northern Poland to western Norway, through the centres of final ice sheet decay in Norrbotten, is show in Fig. 10a-c using the adjusted varve timescale based on the correlation with GRIP by Andren *et al.* (1999).

3.3 Glaciological implications – streaming within the retreating ice sheet

It has been assumed that, on a relatively flat surface, very strong flow-parallel lineation sets and/or evidence of lobation of the ice sheet margin indicate the fomer existence of an ice stream. However, lobate ice margins can be interpreted in two ways: as the termini of ice streams which protrude beyond the normal ice margin because of high ice velocities along their axes, or as expressions of radial flow from the end of an ice divide or ice dome. In the first case they represent a zone of relatively fast flow, in the second, a zone of relatively slow flow. Lagerlund (1980) has interpreted the southern Baltic Sea lobes as evidence of an ice dome in the southern Baltic rather than of ice streams. Lobes and associated lineations which occur entirely on land, for instance in Finland, show lineations at their margins which are not normal to the margins, as would occur if they were ice domes, but are aligned at angles acute to the margin, which is compatible with an ice stream origin. The fact that evidence for the southern Baltic lobes lies partly beneath the sea does not permit the two hypotheses to be tested as they can be in Finland. It is suggested however that the similarities of form should be taken as evidence of similarity of origin in the absence of strong evidence to the contrary.

The relationship between lobation and strong lineation can be used to infer the time transgressive history of an ice stream and to interpret its length (Boulton *et al.*, 2001). Using this approach, the pattern of time transgressive streaming in the ice sheet during its retreat is shown in Fig. 11. This diagram does not show the length of ice streams at any one stage of their history, but the periods during which ice streams reached the margin of the ice sheet. It shows a lettering scheme which identifies the ice streams on the terrestrial margin of the ice sheet. Ice streams B, F and G are deduced largely from lobation of the contemporary ice margin. Ice stream A is an outcome of the convergence of ice flowing from flanking land areas into the Skagerrak embayment, but there is seismic evidence that it extended along the axis of the Norwegian Channel (Sejrup *et al.*, 2000) and lineations on Jæren, in southwest Norway, where the ice stream transgressed onto land (Sejrup *et al.*, 1997). Others are deduced from locally strong lineation sets.

Lineation evidence for ice streams is restricted to the eastern, northern and southern margins of the retreating ice sheet. By analogy with modern ice sheets, ice streams would also be expected at the marine margins of the ice sheet. Ice stream J for example, is well marked by strong lineations on land and trends distally towards a well-defined, over-deepened shelf trough and a lobate glacial unit, presumed to be a moraine, at the shelf edge (Rokoengen *et al.*, 1979). In a tentative estimate of the locations of former ice streams at the western, marine margin, over-deepened troughs on the shelf and lobate moraines or thick glacial sediment units at the shelf edge (Rokoengen *et al.*, 1979; King *et al.*, 1987; Rokoengen *et al.*, 1988; Rise et al., 1988; Holtedahl, 1993) have been used as possible indicators of the palaeo-ice streams, marked on Fig. 11.

If lobation of the contemporary ice sheet margin, based on moraine distributions (e.g. Chebotareva, 1977) as an index of streaming in the south eastern flank of the ice sheet south of the Baltic and the Gulf of Finland is used, many seem to correspond with streams on the Shield which are clearly indicated by lineations patterns. They have therefore been identified, using suffixes, as ice stream complexes which appear to recur along the same axis and which are often succeeded by single major ice streams at a later stage of retreat. They are:

B1 - B5- Southern Baltic ice stream complex.
E - E1- Karelian ice stream complex.
G1 - G4- White Sea ice stream complex.

3.4 Glaciological implications – the phases of retreat

The structural evolution of the ice sheet during its decay is characterised by sub-dividing it into a series of phases as shown in Figs 8 and 10.

Phase R1 – 29 ka to 20 ka

The growth of the Northern European ice sheet culminated in its southwest, west and northwesterly sectors during this period, when it extended onto continental in the North Sea, the northwestern and northern. It then retreated from these shelves before the GRIP core indicates that a significant climatic warming had occurred.

Phase R2 – 20 ka to 15.2 ka

This is characterised by:

a) a large net retreat in the southern sector whilst the eastern sector was stationary at, or advancing to, the maximum extent;

b) major readvances separated by intervening retreats of the ice-sheet margin in the southwestern and southern sectors.

Phase R3 - 15.2 – 13 ka

This was characterised by general retreat around the whole of the ice-sheet margin, reflecting a strong climatic amelioration. The corrected varve record (Fig. 10a) shows a rapid early retreat followed by a decceleration towards the Younger Dryas moraines. Retreat at the western margin was much slower (Fig. 10b-c). Particularly dramatic rates of retreat occurred in the southern Baltic and White Sea regions (Fig. 8), which were probably enhanced by iceberg calving.

Phase R4 - 13 ka to 11.5/11.6 ka

The Younger Dryas Stadial, which occurred between about 12,860-11,640 yr BP, was associated with a sudden temperature drop of 7°C at the beginning of the period and a similar increase at the end (Johnsen et al., 1992; Bond et al., 1993). Large terminal moraines, the Fennoscandian moraines, were generated in Norway (Ra Moraines), Sweden (Middle-Swedish Moraines) and Finland (Salpausselkä I-III) with smaller moraines in Russian Karelia (Rugozero and Nyukozero Moraines - Punkari, 1985), although only discontinuous moraines were produced on the Kola Peninsula (Punkari, 1993). The ice margin readvanced in western Norway (Mangerud et al., 1979) and western Sweden between 12.8 and 12 ka, although it was either stationary or retreated slightly further east during the same time period.

Phase R5 - 11.5/6 ka to final retreat

Rapid retreat of the southeastern flank ice sheet was resumed and the western, southwestern and northern flanks withdrew from the marine margin more slowly as a consequence of substantial climatic amelioration. After 10.6 ka, the ice sheet broke up into separate domes, the main ice mass centred to the east of the modern ice cap of Svartisen, along the Norwegian-Swedish border, a dome centred over the Jotunheimen area in southwest Norway, and a small dome separating later in central Norway/ Sweden (Jämtland). The last ice caps were located further to the east than the initial growth centres. Some strong lineations entering into the northern extremity of the Gulf of Bothnia from the northwest (Fig. 4a) suggest that some streaming continued here at least into the early part of this phase, although it is not clear whether this belongs to a successor of streams E, D or C, or a newly-created stream.

4.0 OLDER FLOW SETS - THE PATTERN OF ICE SHEET GROWTH

4.1 The pattern of pre-retreat lineations

Crossing lineations have been observed in many areas, where flow-parallel lineations on satellite images and drumlins and striations observed on the ground clearly predate the flow features produced near to the retreating ice margin during final deglaciation (eg. Figs 3-5). Superimposed drumlins reflecting different flow directions were first identified by Rose & Letzer (1977). Boulton (1987) showed how they could be produced by changing ice-sheet flow directions, and Boulton & Clark (1990a, b) mapped crossing lineations which reflect such features over wide areas of Canada from satellite imagery. In Fig. 12, all lineations which lie parallel to the flow during final retreat have been removed from the total lineation population to reveal lineations which unequivocally pre-date final retreat. This is not to say that many lineations which happen to be parallel to marginal flow during final retreat do not pre-date that retreat, merely that our conservative approach does not recognise them. The residuals come predominantly from the areas of crossing lineations shown in Figs 4-5. They tend to form well-defined clusters, which are particularly strong in northern Fennoscandia, although residuals occur throughout central and southern Sweden, in particular in southwestern Sweden. In northern Finland/Sweden and in central Sweden (Fig. 12), these early lineations tend to form two sets, with the earlier set in the more northerly area oriented NW-SE and the later set oriented NNW-SSE. It is suggested that many residual lineations were protected from overprinting by the re-organisation of ice-sheet dynamics during the latest phase of glaciation, by temporary phases of basal freezing (Kleman, 1990), migration of the ice divide over the site (Boulton & Clark, 1990) or because they lay beneath zones of sluggish flow between ice streams when the area was in the near-terminal zone during deglaciation (Fig. 4b).

The pattern of residual lineations, shown in Fig. 12 could have formed far from the ice-sheet margin during the final retreat, or during an earlier stage of ice-sheet retreat or advance, either in the Weichselian or during an earlier glacial stage (Tanner, 1915; Punkari, 1985; Nordkalott Project, 1986).

In southwestern Sweden, the residuals show a north-south trend compared with the NE-SW trend during deglaciation. In central Sweden they show a NW-SE trend compared with a NNW-SSE trend during deglaciation. In northeastern Sweden, northern Finland and northwestern Russia they show a NW-SE to NNW-SSE trend compared to a dominant W-E trend during deglaciation; and in southern Finland they show a N-S trend compared with a NW-SE trend during deglaciation. In the area north of the Gulf of Bothnia, the pattern of residuals is more complex, progressively swinging round towards a north-south trend further to the south and east (Fig. 11). Studies of their interrelations suggest that the NW-SE lineation is the earlier, over-printed by a later lineation which fans out from northernmost Sweden in an arc from SW and W.

The pattern of final retreat on the SE flank of the ice sheet east of the Baltic Sea, with a margin oriented in a NE-SW direction, indicates that the dominant flow during retreat in this sector was NW-SE, from a principle ice divide oriented SW-NE. It is concluded therefore that neither the N-S orientation in southern Finland, nor the earlier NW-SE lineations in northeastern Sweden belong to the final decay phase. However, in the Gulf of Finland region, some of the older N and NE directions may represent flow far inside the receding margin during the last deglaciation, produced by convergent drawdown at the head of the Baltic Sea ice stream.

4.2 Reconstructing the pattern of ice sheet growth

The change in lineation trend between early and late lineations must reflect a shift in the pattern of ice flow. There is relative continuity and integration in the residual lineation pattern (Fig. 12) (apart from in southern Sweden, where a NW-SE lineation is replaced by a superimposed N-S lineation further south, and northern Finland / Sweden, where a NW-SE lineation is succeeded to the south by a N-S lineation superimposed upon it). There is also a systematic ice-sheet wide directional shift between pre-retreat and retreat patterns; clockwise in the southeast and central sector and anti-clockwise in the eastern and northeastern sectors. These observations support that the early lineation pattern resulted from a single glacial phase in which the overall patterns of flow and ice divide distribution were different from those during retreat.

Whilst it is possible that the different patterns were the product of different glacial periods in which the ice sheets had different geometries, Boulton *et al.* (2001) suggest it to be most likely that the early lineations reflect growth phases of the Weichselian ice sheet in which the pattern of flow was different from the pattern of flow during retreat. They also suggest that, apart from any possible streaming (see below), strong lineations able to survive a succeeding glacial phase with a different flow direction are most likely to be produced in the near-margin high velocity zone. If the residual lineations shown in Fig. 12 belong to the phase of ice sheet growth, the general time-transgressive trajectory of the expanding ice sheet margin can be determined (Fig. 12). This pattern is not surprising, for the following reasons:

- The advance along the land sector from northern Finland to the Gulf of Finland appears to have been faster than from central Sweden over the central Baltic Sea. This could reflect slowing of advance beyond the east coast of Sweden because of calving into contemporary deep water in the Gulf of Bothnia.

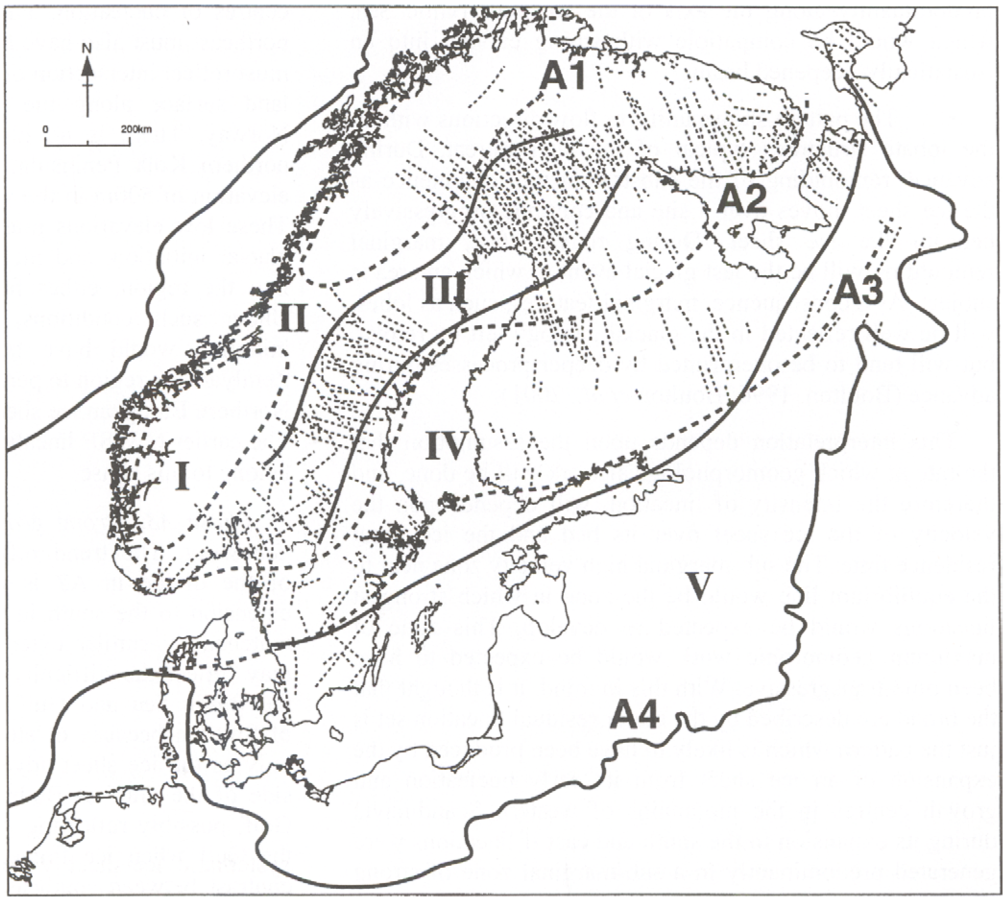

Fig. 12. The major pre-maximum lineation sets which differ in orientation from the retreat sets. They tend to be preserved in areas which lay beneath the principal ice divide at the maximum of glaciation, which protected them from erosion because of frozen bed conditions and low ice velocities (Boulton et al., 2001), and beneath inter-stream ridges (Figs 5, 7, 11) characterised by low ice velocities and possibly frozen bed conditions.

Assumed basal thermal regime. The solid lines identify different lineation zones, and are explained by reference to the time dependent thermal zonation (Boulton et al.,2001). Zone I: no major pre-maximum lineation sets; the area in which there was no basal melting through much of the glacial cycle. Zone II: a single major set of pre-maximum lineations; a zone in which there was basal melting during the Early Weichselian prior to 80 ka, but which was subsequently characterised by basal freezing. Zone III: two major sets of pre-maximum lineations; explained as a zone in which the same Early Weichselian lineation set as those in zone II is crossed by a lineation set formed in the zone of melting during the Middle Weichselian and in the phase of ice sheet growth prior to 25 ka. Its outer limit shows the limit of an Early Weichselian advance. Zone IV: only contains one major pre-maximum lineation set; explained as forming during ice sheet growth immediately prior to the last glacial maximum with a post-maximum lineation being superimposed on it after an intervening phase of basal freezing. Zone V: no major pre-maximum lineation set, explained as a zone which only experienced basal melting, so that lineations have been progressively modified to leave only one major lineation set.

Patterns of ice sheet growth. Pre-maximum lineation sets are used to determine patterns of growth of the last ice sheet based on the assumption that during ice advance the maximum rate of erosion/deposition will be in the terminal zone of the ice sheet (Boulton et al., 2001). It is also based on the view that on a horizontal surface, ice margin trends will lie normal to the flow direction. The general pattern of growth is illustrated by concentric dashed lines, which are arbitrarily used to subdivide growth into four main phases (A1-A4). Patterns of growth in western Norway are entirely speculative, as is the connection between phase 3 and the extension of the ice sheet into the central North Sea.

- The lineation pattern in northwestern Russia and northern Scandinavia suggests that an ice divide extended to the north east at least as far north as the northern coast, and that ice must have transgressed into the shallow waters of southern Barents Sea. This pattern of extension is more likely during an ice sheet growth stage when relative sea levels were low, compared to a retreat stage when they are high because of isostatic depression of the crust.
- A similar argument can be employed in southern Sweden, where it is reasonable to expect an advance into a

shallow early-glacial southern Baltic Sea from the mountainous area to the north. The retreat takes place however predominantly along the axis of the southern Baltic Sea, which would be compatible with strong calving into an isostatically-deepened basin.

- The pattern shows uniform flow directions without the lobate structures typical of the later retreat. During advance, remoulding of the substratum will commence as the ice sheet moves over a site and continue progressively beneath the ice sheet. During retreat, fast, marginal remoulding will be the last glacial event to which an area is subject. As a consequence, marginal features, such as lobes, will be well reflected in the glacial geology left by retreat, but will tend to be overprinted by deeper processes during advance (Boulton, 1996; Boulton *et al.*, 2001).

This interpretation depends upon the assumption that the rate at which geomorphological work will be done, and therefore the intensity of lineation, will depend upon the velocity of the ice sheet over its bed and the ice sheet residence time. The sub-marginal high velocity zone near to the equilibrium line would be the zone in which strongest lineations would be expected to develop. This zone of maximum geomorphic work would be expected to have been time-transgressive. With this in mind, it is thought that the broad arc described by the major residual lineation set is just the pattern which is likely to have been produced by the expansion of an ice sheet from its early nucleation and growth centres in the mountains of western Scandinavia during its expansion to the south and east if lineations were generated predominantly in a sub-marginal zone of strong geomorphic activity.

We speculatively suggest a simple glaciological rationale for the general pattern of ice sheet expansion suggested in Fig. 12 through a subdivision into four phases, though without any chronological implications except in the final, A4, phase.

Phase A1. *Ice-sheet initiation* begins as separate domes in the areas of Jotunheimen/Hardangervidda in southwest Norway and in the Scandinavian mountains north of Kebnekaise in northern Sweden/Lapland, from which they grew to coalesce. Ice-sheet initiation was a consequence of climatic deterioration but the location of nucleation centres was determined by topography. The two early growth centres, in south west Norway and northern Sweden/ Lapland, are currently the two highest and most heavily-glaciated areas along the Scandinavian mountain chain. These are the obvious areas for ice-sheet nucleation, although further growth is not merely depend upon the presence of high mountains, but also the existence of a large high altitude hinterland on which large ice domes can grow. The northern growth centre lay significantly further north than the northern centre during final decay, whereas the southern growth and decay centres seem to have been coincident.

Phase A2. *Radial expansion of the coalesced ice sheet and elongation of the ice divide towards the north and northeast and into the Kola Peninsula.* This expansion was driven by climate and occurred radially away from the centres of nucleation. The extension of the divide to the northeast must also have been climatically-determined, and must reflect intersection of the permanent snowline with the land surface along the Barents Sea coast of northern Norway. There is no high mountain topography in the northern Kola Peninsula. Coastal hills have a maximum elevation of 500m in the west and up to 350m further east. These low elevations must have been adequate to permit glacial initiation, and must reflect a strong moisture flux into the region, either from the northwest or southeast. Under such conditions, it is probable that ice-sheet initiation would have occurred in the Barents/Novaya Zemlya/Kara region to permit coalescence with the growing Northern European ice sheet in the region by late phase A2. The earlier NW-SE lineations in southern Sweden probably belong to this phase.

Phase A3. *Strong growth in the eastern sector of the ice sheet.* The trend reflected by the eastern extension of the divide in A2 is continued during phase A3 by extension to the south in the eastern sector. This was not matched by similar extension further west. This contrast may reflect the difficulty the ice sheet found in crossing the Baltic Sea and Gulf of Bothnia from the east coast of Sweden because of strong calving in deep water. It is only as the ice sheet advanced southwards on the eastern side of the Gulf of Bothnia that ice extended across the Gulf, possibly reflecting greater effectiveness in spanning the Gulf when ice flowed into it from both sides. This contrast between southeast and southwestern sectors may therefore be a dynamic rather than a climatic effect. The apparent failure of the ice sheet to cross from southern Sweden to Denmark in an area lacking large expanses of deep water might however indicate a strong climatic gradient between a cold but still moist eastern area, and a much warmer western area. It is presumed that the ice sheet advanced in the west at least as far as the coastal zone during phase A3 or even A2. The strong growth in the eastern sector and limited growth in the west during this phase implies a swinging of the ice divide about a pivot in southwest Norway, from a SW-NE orientation during phase 2 to a WSW-ENE orientation by the end of phase A3.

Phase A4. *Expansion in the central southern sector.* The ice sheet extended to, or close to the south west coast of Norway by the end of phase A3, and then expanded into the central North Sea at about 28 ka (Sejrup *et al.,* 1994) during A4, when it also expanded in the central southern sector. (Phase A4 and phase R1 overlap. The ice sheet dome over the central North Sea appears to have collapsed before the ice sheet reached its maximum extent in the southern and southeastern sectors.) This expansion would have been associated with a bowing towards the south of the primary ice divide, or the development of a secondary divide towards the south . This could account for some of the SW-directed lineations in the Stockholm area.

Further east, in the Baltic Sea and Finland, there was a very large advance, compared with further west in the North Sea area, where the ice sheet had already reached its maximum advance by early in A4, and further east in northwest Russia. This is consistent with recent AMS ^{14}C re-dating of mammoth remains (Ukkonen *et al.,* 1999) from six sites pre-dating the last, Late Weichselian ice sheet advance. These have produced ages ranging from 31,970 ± 950 BP from northwestern Finland, immediately west of the head of the Gulf of Bothnia, to 22,420 ± 315 BP in central Finland to 23,340 ± 350 BP at Helsinki in southern Finland. Notwithstanding the uncertainties associated with the dating of bone of this age (e.g.: Berglund *et al.,*1976), the evidence suggests that the ice sheet advanced by some 700-800 km to its maximum extent after about 22-23 ka. It is also consistent with the evidence from northwest Russia of a late advance to and retreat from the regional maximum, although the extent of the A4 advance there is less than further west. The N-S lineations in southern Sweden which post-date NW-SE lineations belong to this phase.

The moisture sources which fed the expansion in northwest Russia in phase A3 may have been cut off from the westerly moisture sources by ice-sheet growth. Further west, the central southern sector expansion may have reflected increasingly maritime conditions, which in the extreme west had become so warm as to inhibit further growth. Such changes may be explained by some of the relationships between storm tracks, maritime/continental contrasts and glacierisation pointed out by Kvasov (1978). The same highly maritime conditions in the south west would have created a relatively energetic ice sheet, which was able to cross the deep waters of the Norwegian Channel (300m depth at the present day) and extend into the central North Sea region to become confluent with a British ice sheet, when we assume that a primary ice divide would have extended from south west Norway into the central Norh Sea. It is thought that advance to the continental shelf edge in the west occured during phase A4.

Evidence from the Kattegat area of south Sweden/Denmark suggests two early stage ice advances (Sjørring, 1983; Houmark-Nielsen, 1987) a Norwegian advance and an Old Baltic advance. Sjørring (1983) regarded the former as the earlier of the two, whilst Houmark-Nielsen (1987) and Ringberg (1988) believed it to be the latter. The succeeding Main Danish Advance and later readvances are believed (Houmark-Nielsen, 1987) to have coincided with the LGM and early retreat stages from it. The authors suggest that advance from the north in the area of Jutland could belong to late phase A2, A3 or early phase A4 conditions, whilst in Sweden they are most likely to reflect A4 conditions.

The absence of evidence of flow from a Barents Sea ice mass in the NNE and from a Kara Sea mass in the NE, suggests that ice on the European mainland was dominantly Europe-centred and can largely be considered as a single dynamic entity.

4.3 The chronology of ice sheet fluctuation prior to the Last Glacial Maximum

There are currently two schools of thought about the tempo of glaciation in northern Fennoscandia. One group (Group A: e.g.: Korpela, 1969; Hirvas & Nenonen, 1985; Lagerbäck, 1988; Kleman, 1990; Sutinen, 1992) argues for the preservation of tills, eskers and major end moraines from the early Weichselian and Saalian glaciations, which were separated from the phase of advance to the Weichselain maximum by periods of deglaciation. This is based on discoveries of organic matter underlying the uppermost till unit and which are interpreted as evidence of deglaciation. They suggest that the organic units separate two glacial phases during which the patterns of ice flow (determined from till fabric) were quite different. Another group (Group B: Aario & Forsström, 1979; Punkari, 1984, 1993; Forsström & Eronen, 1991; Punkari & Forsström, 1995) argues that much of the organic material is redeposited, is not therefore evidence of deglaciation, and that the inter-till organic beds are seldom found between tills reflecting the two different flow patterns in question.

The views of group A imply strong ice-marginal oscillations through the Weichselian, with the ice-sheet margin withdrawing to northern Fennoscandia during interstadials (Mangerud, 1991; Lundqvist, 1986). Group B stress a longer, slower and more sustained ice sheet build up to its maximum Weichselian extent, with less dramatic pre-LGM oscillations on the eastern flank of the ice sheet.

Boulton et al. (2001) have used a glaciological flowline model to simulate the glaciological implications of the view of group A. It generates two well-integrated sets of time-transgressive glacial lineation. The earlier set would is best interpreted as a combination of sets formed in zone II, at about 90 ka, and in zone III, between about 50 ka and 20 ka, with the younger set representing deglaciation in zone III.

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