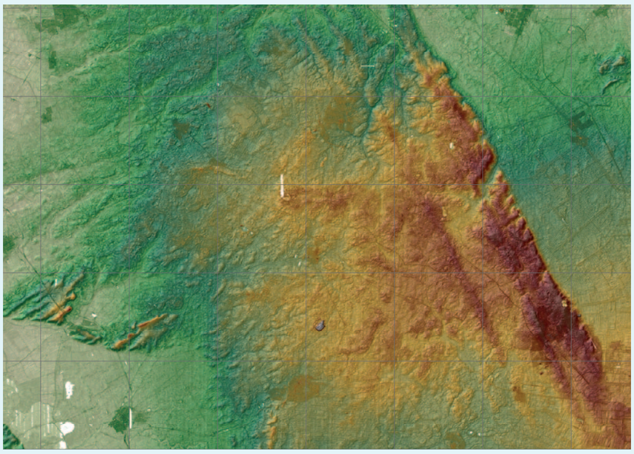
An integrated approach to reconstruct the Saalian glaciation

GIS-based construction of a new phase model for the Netherlands and NW-Germany



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On the front page:
Above: ice field with meltwater pools
Bottom left: Deformation of Tertiary strata in till section De Lutte (from: Rappol, 1993)
Bottom right: AHN image of the province of Drenthe

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Abstract

Some 170,000-150,000 years ago (during MIS 6), large ice masses last covered the Netherlands and NW Germany (Saalian Drenthe Substage). This left many geomorphological features in the landscape, e.g. ice-pushed ridges, sandurs and glacial basins. Throughout the 20th century extensive research has been revealed on this geomorphological assemblage and produced interpreted sequence of glacial events, known as glaciation phase models. The successive competing phase models of the 1960ties to the 1990ies each appear biased to specific features, subregions and types-of-data. At present, new data and insights exist that have risen since the construction of the currently established models.
In this research the sequence of events was newly reconstructed, aiming to unify the evidence in NW Germany with that in the Netherlands. I collected geological-geomorphological evidence (literature inventory) and newly interpreted high-resolution elevation data in an inventory GIS. The conceptual phase models and related glaciological processes during the glaciation were reviewed, responsible for the eventual ice-margin landscape. Elements of 'classic' knowledge and new lines of reasoning are each outlined. The newly constructed phase model recognises three phases towards maximum ice-sheet extent, one transition phase and two deglaciation phases. The GIS stores the preferred phase model, as well as earlier interpretations.

1. Introduction

Near the end of the Saalian stage (MIS 6; figure 1.1) large ice masses covered the Netherlands and NW Germany for the last time in geological history (figure 1.1). This glaciation, known as the Drenthe Substage, left many geomorphological features in the landscape, amongst others ice-pushed ridges, sandur plains and glacial basins. Throughout the 20th century extensive research has been done on this geomorphological assemblage and the sequence of glacial events that produced it. The result of that research is a series of phase models of the glaciation. Each of the phase models each appeared biased to specific features, subregions and types-of-data. Also, they are on aspects incompatible between neighbouring regions where other types of features exist that were studied in different ways. At present, new data and insights about glaciations in NW Europe as well as glaciological processes have risen since the early nineties when the most recent phase model for the study area was constructed. Therefore, an update of both the dataset of Saalian features and knowledge on their processes of formation (phase model) was required.

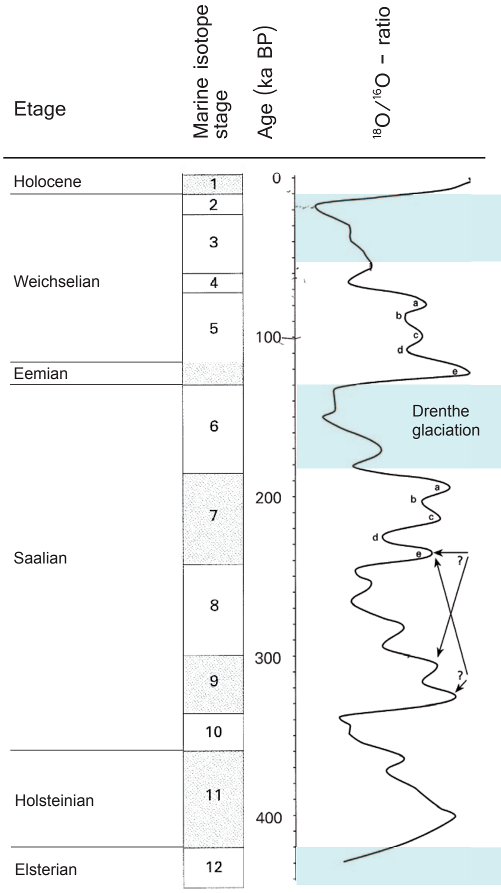


Figure 1.1
Time-line of the Middle and Late Pleistocene. The blue periods mark the coldest periods in which large ice masses certainly reached the research area. The Drenthe glaciation took place in MIS 6. (after: De Mulder et al., 2003)

The research area

The research area is located on the south western margin of the large Late-Saalian Fennoscandian ice sheet (figure 1.2). Most studies have been done on either or the other side of the border. This had yielded two different sets of literature with a deviating content. For understanding the sequence of the glaciation in this area however, it is crucial that these studies are integrated. Therefore, data from the North Sea area, the Netherlands and NW Germany are integrated in this research (figure 1.3).

The aim and use of the research

The aim of this research is to newly reconstruct the sequence of events in the research area and to yield a better understanding of its glacial morphology and geology. This is done by reviewing the Saalian glacial morphology in the research area and assembling the observational evidence into a GIS dataset. It includes glacial, proglacial and deglacial features. The work provides an integrated overview of the knowledge on Saalian glaciation history. Both the 'classical' knowledge and new lines of reasoning are outlined. The compiled GIS dataset presents the knowledge in an accessible and easy manageable integrated form. Once digitized and structured, the study used the GIS as an environment to (re) compile and interpret phase model reconstructions. The improved structuring of data and insight regarding the phasing of the glaciation, contributes to better understanding of the glacial morphology and geology of the area.

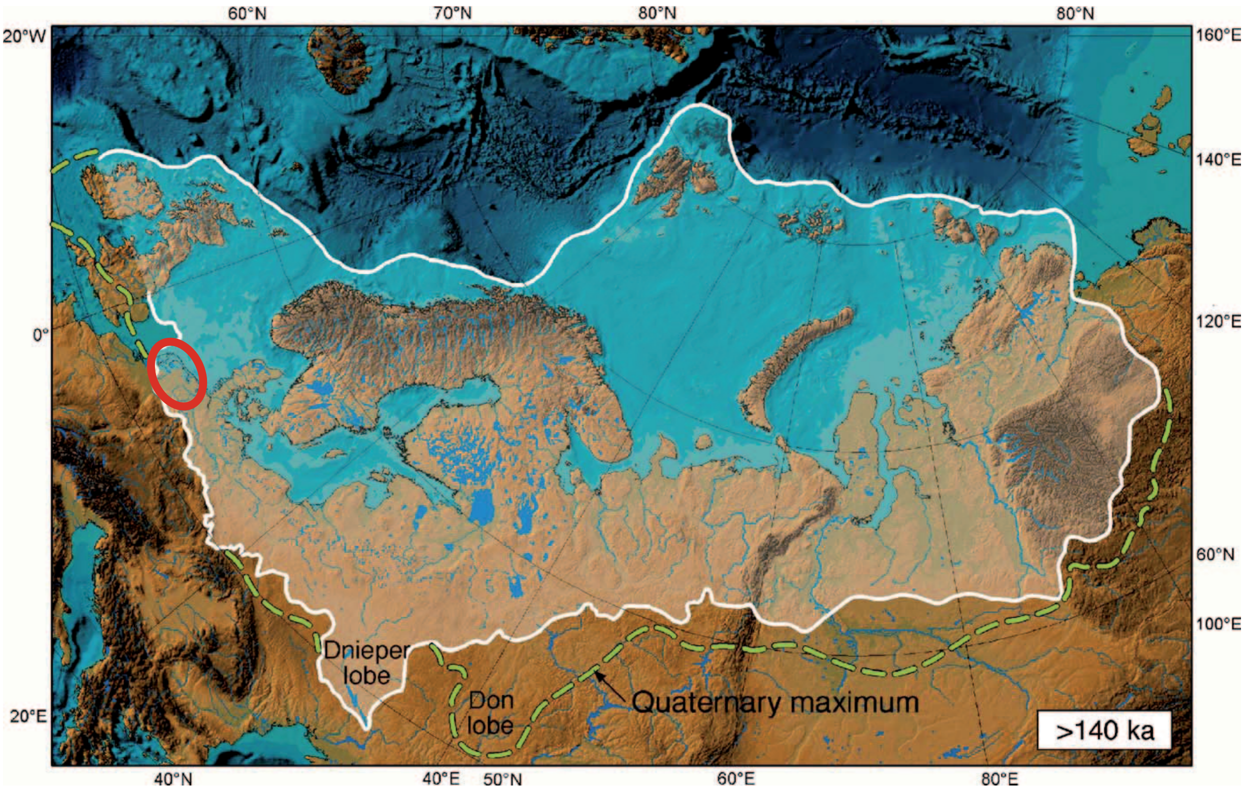


Figure 1.2
The extension of the whole Fennoscandinavian ice sheet during the Drenthe stage. The research area is marked with a red circle. (from: Svendsen et al., 2004)

Updating the phase model of the glaciation of the Drenthe Substage glaciation is required to serve as a framework for studies concerning the Saalian in this region and to use in today's practice with high density datasets (information age). Besides, challenging environmental questions demand up-to-date information on the landscape and subsurface. This study provides regional context for the PhD thesis of Enno Bregman (province of Drenthe), which focuses on the subsoil of the province of Drenthe and its effect on current ecological and hydrological processes in the region.

Thesis outline

The outline of this report is as follows: Chapter Two explains the approach (strategy, project outline) of this project in detail and the structure of the GIS. Chapters Three, Four and Five each are review chapters. Chapter Three gives an overview of the research area and its geological history prior to the Saalian glaciation. Chapter Four gives brief physical background on glacier dynamics. The genesis of landforms and deposits resulting from the ice sheet in the study area are described in chapter Five. Chapter Six and Seven give an overview of classical data integrated with new results, both chapters are richly illustrated with maps derived from the GIS. In Chapter Six an overview is given on the research on glacial geomorphology and deposits in the research area. In Chapter Seven some models on glaciation sequence in the area are outlined and discussed. Finally, in Chapter Eight the new phase model will be presented, again illustrated with figures derived from the GIS.

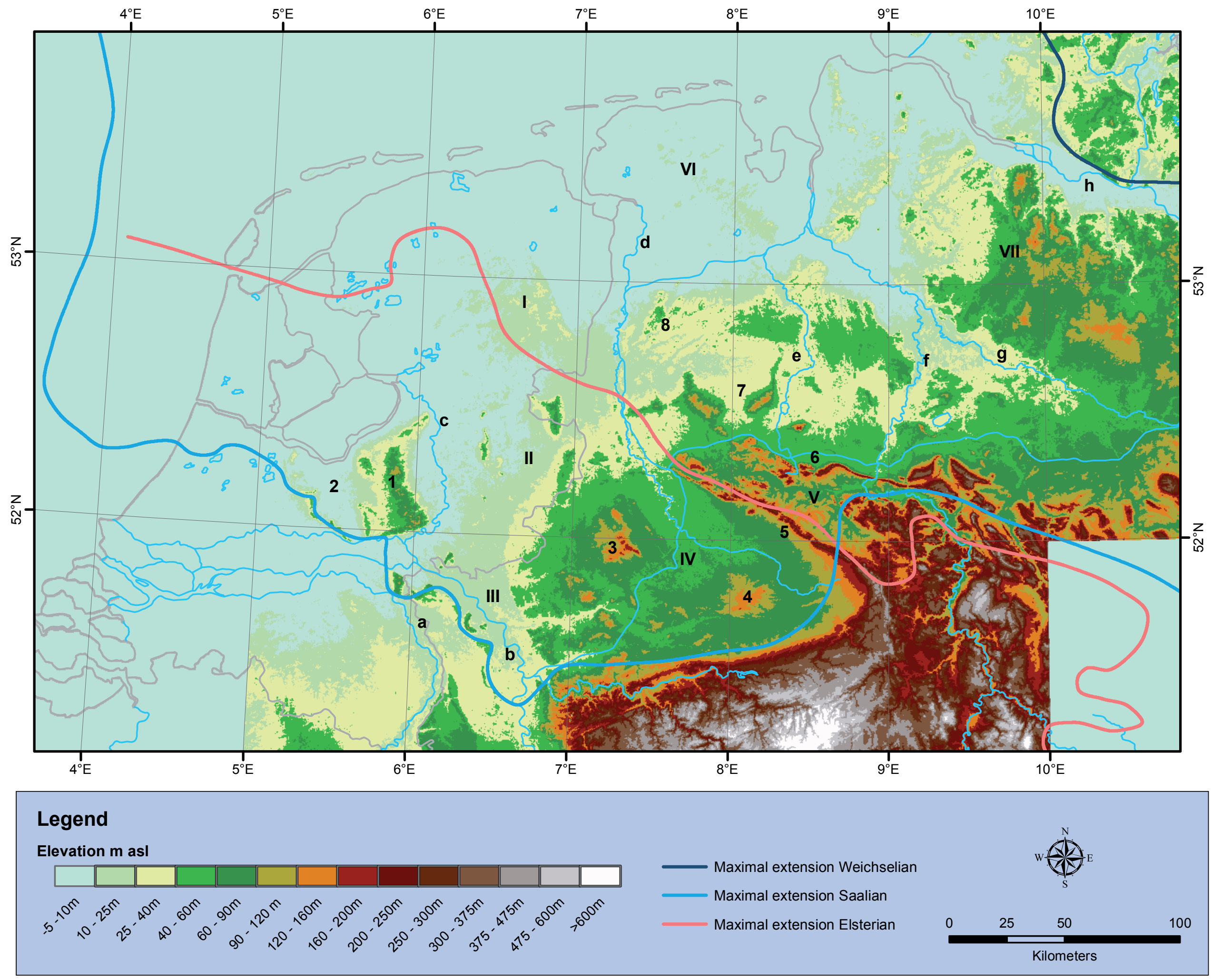


Figure 1.3
The research area covers the southern part of the North Sea, the Netherlands and northwestern Germany. Several regions are indicated on the map: I) the province of Drenthe, II) Twente, III) Lower Rhine embayment, IV) Münsterland embayment/Münsterbasin, V) Weserbergland, VI) Ostfriesland, VII) Lüneburger Heide, 1) Veluwe, 2) Gelderse Vallei, 3) Baumberge, 4) Beckumerberge 5) Teutoburgerwald, 6) Wiehengebirge, 7) Dammer Berge (Rehburg line). The main rivers are: a) Meuse, b) Rhine, c) IJssel, d) Ems, e) Hunte, f) Weser, g) Aller, h) Elbe. The maximal extensions of the last three glaciations are indicated, backgound image: satellite SRTM.

2. Approach and Methods

2.1 Construction of the glaciation phase model

The goal of constructing a phase model for the Drenthe Substage glaciation is to present and describe a sequence of events in correct order for a large area. These 'events' include flow direction, stages of certain maintained positions of the ice front, presence of ice-marginal lakes, activity of river valleys, the formation of geomorphological features (erosional and depositional) and the deposition of sediments. A phase is defined as a stage of distinctive ice flow towards the ice margin that left significant geomorphological evidence. Between localities the signature of the ice flow may slightly change and may be somewhat diachronic, but still be considered to represent one and the same phase. This is the case when a series of neighbouring ice-pushed ridge complexes line up (Rehburg line, Haarlem-Utrecht-Nijmegen-Düsseldorf - HUND-line). They did not necessarily form at the same time, but still they are marked as the product of one phase.

To achieve successful construction of the phase model a couple of difficulties have to be overcome. To start with, the availability of data is limited in large parts of the region, for example offshore, in areas that are covered by thick younger sediments (glacial basins) and in areas that suffered major erosion after the Saalian. Another complication is the fact that direct dating methods for this time range have only become available very recently. This means that phase-model proclaimed (a)synchronicity of events, cannot easily be independently verified using chronometry. The numerical dating methods (OSL, predominantly) simply are not accurate enough to individually resolve the events within the ~5,000 to 10,000 years that the phases of the Drenthe glaciation span. Therefore, relative dating remains the main tool for establishing chronology of glaciation events. Relative dating methods, with principles from geomorphology and stratigraphy as their basis, work well at local scale but are difficult to apply at regional scales. This is due to the fact that relative dating is best applied on individual adjacent landforms that truncate each other. At regional scale too many features may be present and clear truncation relations cannot always be determined.

Because the density of data is not that high and because the chronology of local events cannot always easily be coupled to regional phases, the phase model should not contain too many phases. At least the following phases are to be included: the 'before', the onset, the maximal extension and the deglaciation phase. If appropriate, each of these phases can be split up further, for example when radically different orientated ice flows occur or when a distinctive group of ice-pushed ridges was formed that cannot be explained within a certain phase.

While performing the reconstructions, the new and integrated approach were considered at three scales:

1. Large scale dynamics: the whole ice sheet or its southwestern margin ('zoom out').
2. Mesoscale: assemblage of features (lined-up ice-pushed ridges, fields of flutes etc.).
3. Small scale: individual geomorphological features; e.g. from which side were ice-pushed ridges pushed? ('zoom in').

In chapter 6 features are considered on a small scale and on a mesoscale. In chapter 7 and 8 the elements from chapter 6 are scaled up to mesoscale and integrated with large scale dynamics.

In preparation of the phase model compilation, several activities were undertaken. This is the orientation phase in the research scheme (figure 2.1):

1. Available knowledge on glacial landforms and deposits was described in the context of glacial

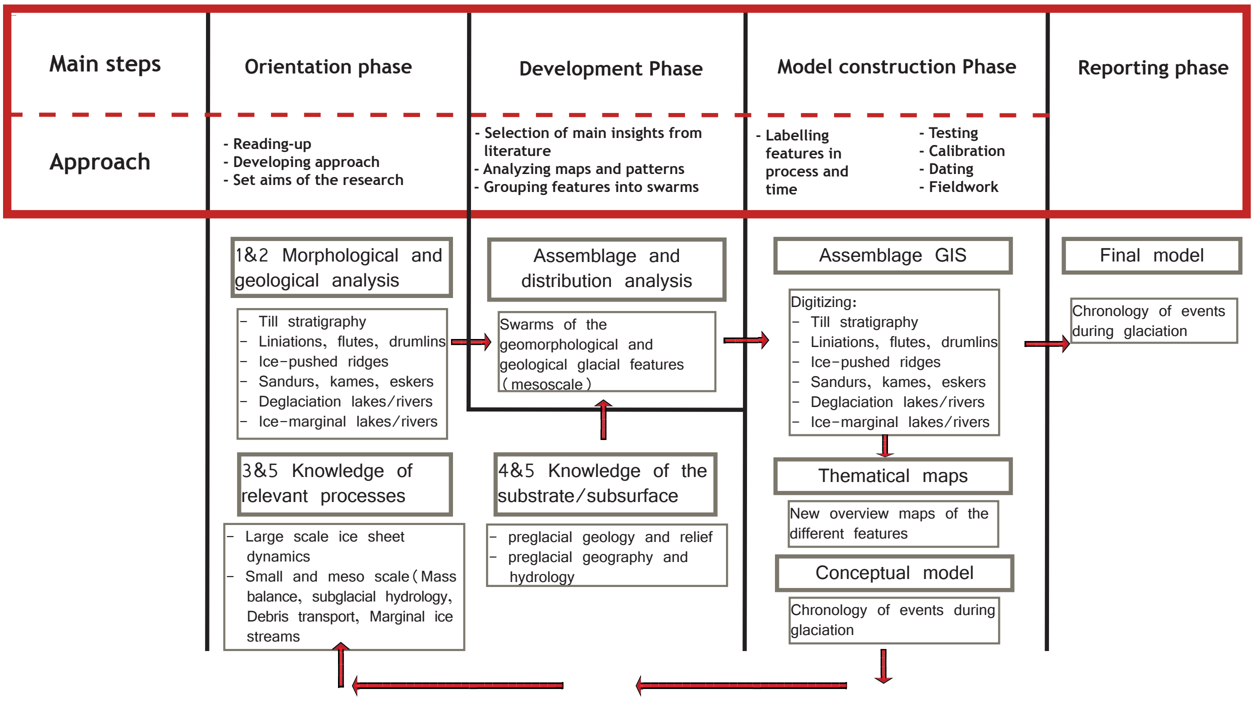


Figure 2.1
Research approach scheme. In the orientation phase, the knowledge (activities 1-5 mentioned above) are collected and integrated in the development phase. Then, the features are assembled in a GIS (model construction), with supporting information from glacial dynamics and substrate knowledge the phase model was deduced. During the reporting phase this information was described in a report. During the construction of the GIS and concept model the integrated knowledge is iteratively checked and improved, which serves as an improved input for the GIS.

dynamics, including the newest insights (small scale and mesoscale).
2. Besides glacial morphology and ice sheet dynamics, also glaciofluvial and glaciolacustrine 'proglacial' data has been integrated (small scale and mesoscale).
3. Ice sheet dynamics of the Scandinavian ice sheet were linked to the ice front movements towards the ice sheet margin of the southwestern sector (large scale).
4. A basic idea on the landscape is outlined (position of valleys, position uplands) and substrate composition, prior to the glaciation (mesoscale).
5. Aspects of glacial dynamics, subglacial hydrology and thermal interaction with the deeper substrate are taken into consideration (mesoscale, small scale).

Activities 1 & 2 include the observational data that was integrated in this study and are the main focus of the research. How these landforms can serve for the glacial reconstruction is also outlined in chapter 2.2.
Activities 3 & 4 provided boundary conditions to embed the phase model in.
Activities 3 & 5 use conceptual knowledge on glaciation in addition to observational data

These steps were done in the orientation phase of the research. In the development phase, the relevant data was assembled and finally incorporated into the GIS (figure 2.1). Theoretically, all this knowledge should fit a phase model, explaining the formation without any contradictions. In reality, difficulties can occur; large areas have too little data or data that can be interpreted in too many ways. To cope with this, the facts were separated as much as possible from the interpretations. The different interpretations and datasets from other studies were compared in order to reinterpret the conflicting interpretations. The integrated facts formed the base of the new glaciation model. Below the main focus areas of the study are outlined.

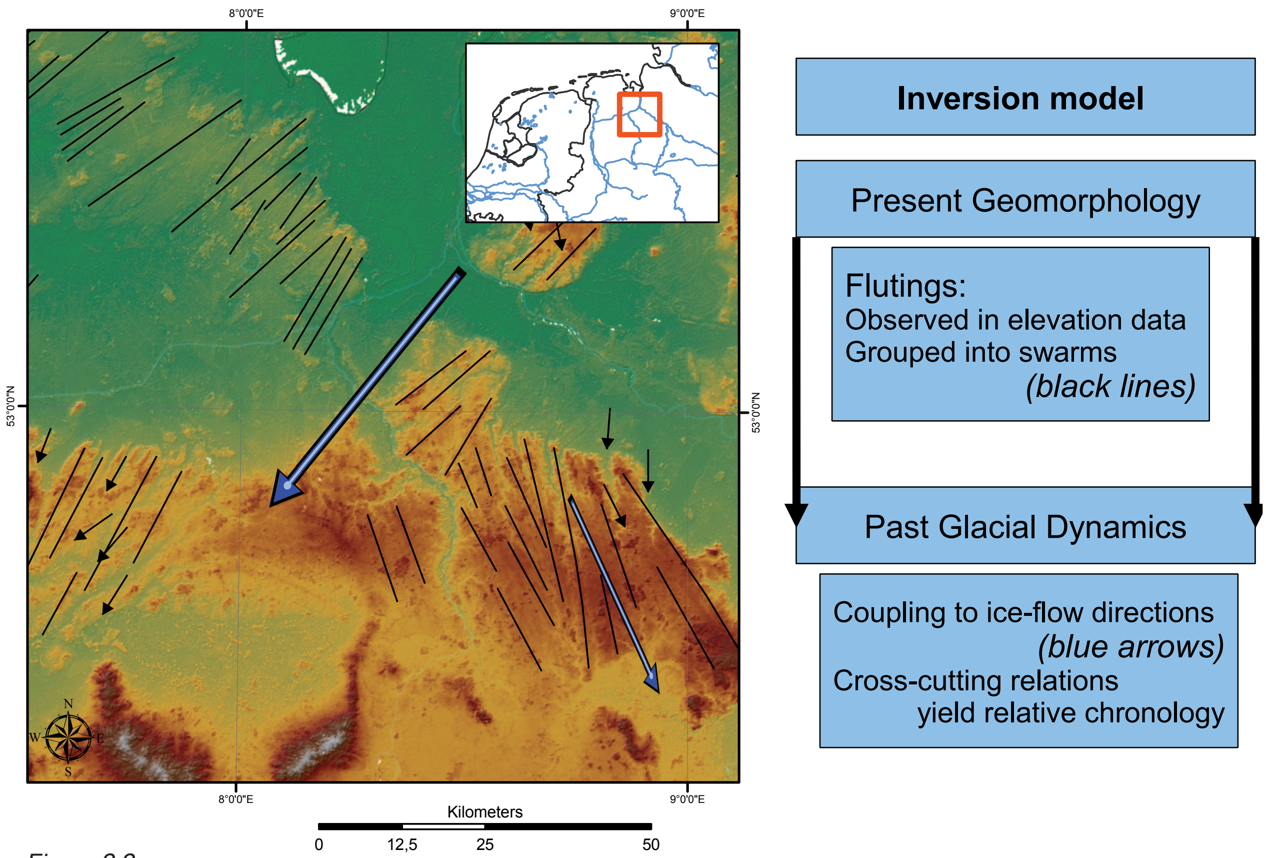


Figure 2.2
The inversion model. Glacial dynamics can be deduced from glacial morphology and its cross-cutting relations.

2.2 Inputs to the phase model

2.2.1 Glacial landforms and deposits

Glacial landforms are traces of past glacial dynamics and can hence be used to reconstruct past conditions of glaciation. This is called the construction of inverse models (cf. Kleman et al., 2006). This thesis contains an outline of general properties of the most relevant glacial landforms and deposits (chapter 5, serving as background information) and a more specific review of the glacial landforms in the research area (chapter 6, description of the observational data). The focus is on tills and till plateaus, glacial lineations and streamlined features, ice-pushed ridges, glacial basins, proglacial lake deposits, ice marginal rivers and deglaciation valleys. Many of these landforms have been investigated by several workers in the last several decades, some more extensively than others. Much of this research was done between 1965 and 1995, after which most pits closed and most of the classical phase models had been constructed (chapter 7). New datasets, especially high resolution digital elevation models (AGI, 2005; SRTM- Farr et al., 2007) have become available since, as have new research techniques, such as ground penetrating radar (GPR e.g. Bakker, 2006). These give partially new insights in the glacial morphology, and partial confirmation of former results.
Furthermore, geological 3D models of the substrate are used (TNO - Digital Geological Model - DGM, 2009). The DGM was used to visualize the already mapped morphological and geological situation from both the top and the bottom of the Saalian deposits. In this way the interaction between the ice and the substrate can be understood better. Besides the morphology of the subsurface and the basal erosional surfaces of the valleys that formed during deglaciation could be visualised. For Germany these models are not yet available. Instead profiles of the Landesamt were used (LBEG, 2006).

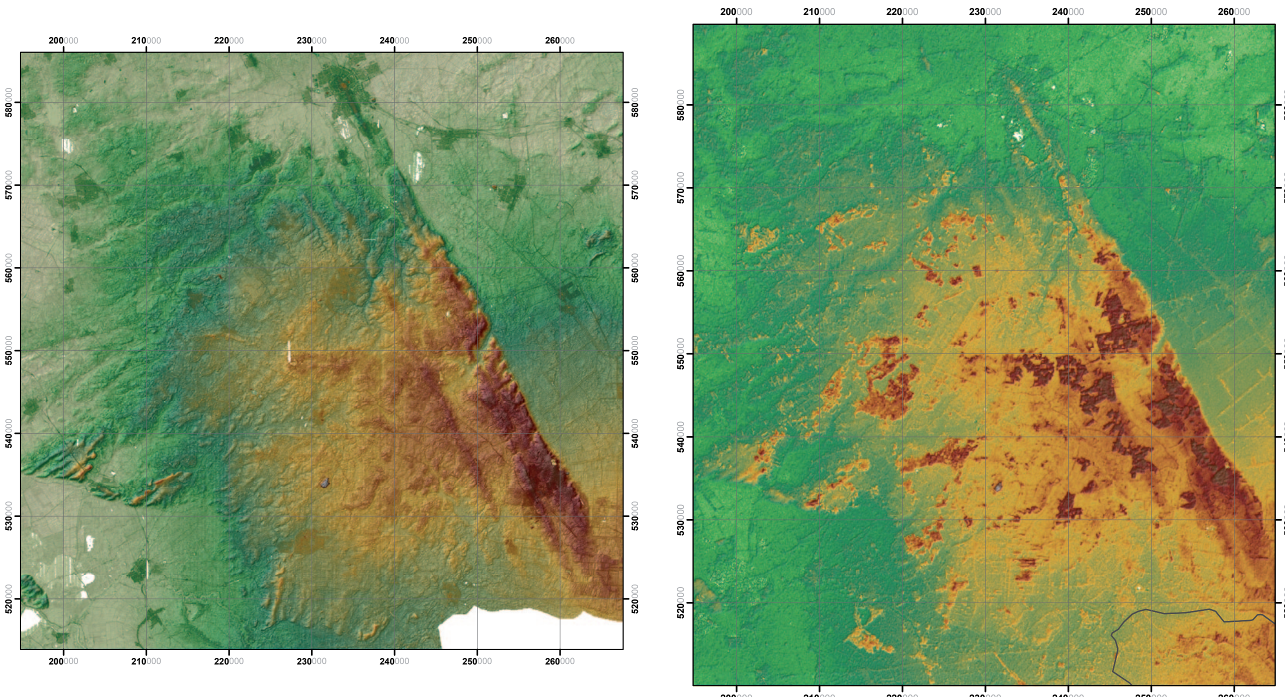


Figure 2.3
The elevation of the province of Drenthe according to the AHN (left) and SRTM (right) datasets. Both give a good impression of the (glacial) morphology of the area. AHN has canopy effects removed and it has the higher resolution and accuracy, consequently more detailed glacial lineations can be seen.

The use of DEMs and 3D models for phase model reconstructions is a new application of these datasets, applied for the first time in this research area and combined with the 'classical' methodologies. DEM-based glaciomorphological mapping has been used in Scandinavia (Boulton et al., 2001; Winsborrow et al., 2010) and Scotland (Evans et al., 2009). In this research it is applied as a new technique in this area. High resolution Digital Elevation Models (DEMs) of the area were used on which ice moulded or streamlined subglacial features can be visualized very clearly. These structures include flutes, megaflutes and drumlins and they reveal the paleo ice flow direction (chapter 6.2.2).
For the Netherlands a high resolution DEM was available: the AHN (Actueel Hoogtebestand Nederland) which has a horizontal resolution of 5 meter and a vertical accuracy of 15 centimetres (resolution = centimeters). It was mapped between 1997 and 2003. For Germany SRTM 3 (Shuttle Radar Topography Mission) was used, it has a 90 m horizontal and 16 m? vertical accuracy (resolution = meters). It was mapped in 2000. Merged visualisations of these datasets were generated in ArcMap (ESRI, v 9.3) and morphologically interpreted. Examples of these elevation images can be seen in figure 2.3. The streamlined features were indicated on these maps. In order to reduce the amount of data they will be reduced by using cartographical representations of coherent directional landforms (mesoscale), called swarms (cf. Kleman et al., 2006). Then, conclusions were drawn regarding the flow direction of the ice in the area (cf. Boulton et al., 2001; Bennett & Glasser, 2009 – chapter 12; Evans et al., 2009; Winsborrow et al., 2010). In some cases several of these features with a different orientation are superposed upon each other indicating different ice flows. When the superposition of features can be deduced from the cross-cutting relations, the relative chronology of different ice flow directions can be obtained (figure 2.2). These can be grouped in stages representing the phases of the glaciation model.

2.2.2 Ice-marginal and deglaciation landforms

Erosional and depositional landforms that formed in the direct vicinity of active ice margins in many cases are directly connected to the ice sheets, as they were generated by melt water released by the ice. This meltwater fed rivers and in the lowest areas their water ponded to form lakes.

These features provide important information on the presence of the ice margin and the melting conditions, which again may have implications for the glacial dynamics. The dimensions of the meltwater systems can give relative age estimations, for example, small river valleys may indicate either low discharge or short developing time. Deposits of ice-marginal rivers may be traced along the ice front, confirming that the ice stagnated for some time at that position. Volumes of sediment trapped as sandurs may be used to estimate the duration of stagnation of the ice front. Also deglacial river valleys are considered, which left significant geomorphological traces in the landscape. The depth of the incision of these channels is indicative for the relative elevation of the erosion base at the time of formation and the amount of sediments it carried. In turn, these observations can be used to order followed-up valleys en sequence; both in proglacial situations and in deglacial situations (chapter 6).
Lake deposits and lake-deltaic deposits may give valuable information on the fluctuation in lake levels, which can again be sequenced and correlated to the ice sheet dynamics. The infilling of lakes that formed during deglaciation can supply information on events during or directly after the deglaciation. During the build-up and maximum extension of ice, ice-marginal lakes and rivers may have influenced the heat balance at the ice margin. The latent heat of these features influenced temperature and hence the groundwater flow under the ice in its surroundings, which co-influenced the ice-flow dynamics towards the ice margin (chapter 4.6).
Last major advantage of including glaciofluvial is the relatively new ability to OSL-date some of the deposits, which indirectly yields a date for the adjacent ice front. Such studies have been performed using proglacial lake deltas and ice marginal river deposits in the research area (e.g. Busschers et al., 2008; Winsemann et al., 2009).

2.3 Embedding the reconstruction

In this paragraph several phenomena and theories that play an important role in the behaviour of ice sheets and ice streams are described, they could not be included in the GIS. This information was used as background information ('embedding' information) for the reconstruction.

2.3.1 Ice sheet dynamics
Knowledge of the large scale glacial dynamics (ice sheet sector scale) is important for reconstructing and explaining the large scale ice flow in the area. It serves as the physical-mechanic context of the phase model. In chapter 4 the large scale dynamics are outlined including the forcing and build-up of ice sheets, mass balance and large scale ice flow, and hydrological dynamics. In the last decades substantial growth of understanding has been reached regarding the ice sheet dynamics of the Scandinavian ice sheet in the last glacials (e.g. Boulton et al., 2001; Svendsen et al., 2004; Mangerud et al., 2004; Van den Berg et al., 2008). This was mainly done using the case of the ice sheet from the last glacial, because the evidence for reconstructing this ice sheet is most abundant. However, many insights from these studies in principle also apply to the analogous penultimate Saalian glaciation (analogy explained in Chapter 4).
At a smaller scale, the marginal ice sheet dynamics are considered that formed the individual glacial landforms (e.g. glacial lineations, drumlins and ice-pushed ridges). To reconstruct how the ice tongues behaved in the area it is important to understand if and how these ice tongue activity was linked to the large-scale glacial dynamics. At mesoscale the occurrence and distribution of glaciogenic geomorphological features in groups is considered (e.g. lines and series of ice-pushed ridges, groups of drumlins). At this scale the geomorphology in the regions can best be linked to ice sheet dynamics. For groups or lines of ice-pushed ridges, for example, hypotheses on the relation with the subsoil and glacier dynamics can be tested in a better way than for individual features. Therefore, some highlights on smaller scale and mesoscale glacial dynamics are included in chapter 4; including warm/cold based glaciers and ice streams & surges. In chapter 5 the glacial landforms and deposits are described and presented in the context of the glacial dynamics of chapter 4. The overviews of chapter 4 and 5 restrict

themselves to the 'glaciated sedimentary basin' situation of the study area.

2.3.2 Antecedent geology: composition of the (deeper) subsurface

The geological history prior to the Saalian and the geological situation of the area at the time the ice arrived define the antecedent conditions for the glaciation. These determined the preglacial surface relief and substrate properties. Preglacial relief is an important boundary conditions for the movement of the ice front, high elevated topography can serve as an obstacle for the ice and therefore it can determine the pathway of large ice flow patterns. The properties of the glacial bed (e.g. coarseness, thickness and heterogeneity of the material) is important for the subglacial hydrology and therefore for ice-bed interaction and ice flow (Van den Berg & Beets, 1987), influencing the glacial morphology at mesoscale. For the Netherlands, the formations that at present occur directly below the Saalian glaciogenic levels, were extracted from the Digital Geological Model (DGM, 2009) – see chapter 3.5.
A correlation seems to exist between structures in the deeper subsurface and the glacial morphology (e.g. Van Balen et al., 2005). The GIS allows to overlay the glacial morphology with other data sources, such as deeper subsurface structures. The results are presented in chapter 6.2 to explore suggested links between the substrate and glacial morphology.

2.3.3 Subglacial hydrology and geothermal processes in the ice-marginal area

Subglacial hydrology is considered to be an important mechanism for the glacier dynamics because it mainly determines the thermal processes underneath the ice sheet (Piotrowski, 2006). This is an important mechanism in glacier dynamics, especially in the marginal regions where the ice is relatively thin and the base wet and warm (Bennett & Glasser, 2009). Here the ice is very sensitive for heat sources, which can trigger and guide ice streams (Winsborrow et al., 2010b - chapter 4.4 and 4.6). Subglacial hydrology is mainly determined by the composition and structure of the shallow and the deeper substrate combined (chapter 3.3). Faults, basin slopes and salt domes may have severely affect deeper groundwater flows and thereby the thermal regime underneath an ice tongue (chapter 4.4). Maps of the temperature of the subsurface are useful and can be combined with overlays from the GIS to explore correlations.

2.4 Assembling the phase model in GIS

The glacial features in the research area were digitized in a GIS. This was done to provide a decent overview and a solid database of the glacial features. This overview is important to visualize unmapped or hardly mapped features and to avoid that very well visible features (e.g. ice-pushed ridges) to become too dominant during the construction of the phase model. The GIS consists of labelled polygons in one single layer containing glacial-geomorphological information. Its setup is such that it yields a balanced overview of the features and that no white spots occur the study area. This forces the researcher to consider the whole region when discriminating between the phases. It is not a GIS that stores observation only – it also stores the interpretations independently. In its setup it is similar to the channel-belt GIS for the Holocene Rhine-Meuse delta (Berendsen et al., 2007).

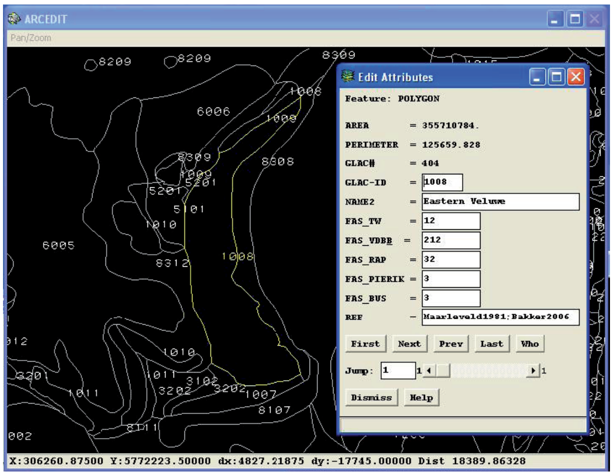


Figure 2.4
Labelling in the GIS. The different polygons in the Veluwe region are digitized and labelled in ARCinfo Workstation.

Features	'Glac-ID'
Push moraines	
extramarginal, (formed during maximal extension)	1001 – 1099
extramarginal, readvance)	11..
intramarginal, overridden	12..
Intramarginal, drumlinized or fluted	13..
Till plateaus	
till plateaus with till deposits	2001 – 0299
till plateaus residual till deposits	22..
eroded till plateaus	23..
residual tills on push moraine	24..
pre-glacial deposits in the till plateau	25..
till plateau on sandur	26..
Sandurs	
intramarginal, overridden	3001 – 3099
extramarginal, (formed during maximal extension)	31..
extramarginal on push moraine	32..
extramarginal, formed during readvance	33..
Eskers	
present as ridges	4001 – 4099
present in tunnel valleys	41..
Extramarginal lacustrine landforms	
extramarginal kames	5001 – 5099
extramarginal kames on push moraine	52..
deglaciation lake (deposits)	54..
ice-marginal lake (with or without much deposits)	55..
Intramarginal lacustrine landforms	
intramarginal kames	6001 – 6099
glacial basin (extramarginal)	60..
glacial basin (extramarginal, readvance)	61..
glacial basin (intramarginal)	62..
Ice marginal river valleys	
river valleys, extramarginal	7001 – 7099
Deglaciation river valleys	
intramarginal, eroding (till/drumlin etc)	7101 – 7199
intramarginal, eroding (extramarginal ice-push/sandr erosion))	72...
intramarginal, eroding (readvance extramarginal ice-push/sandr erosion))	73..
intramarginal, eroding (intramarginal ice-push/sandr erosion))	74..
intramarginal, depositional (basin filling, deglaciation)	75..
intramarginal, eroding mainly presaalian sediments	77..
intramarginal, depositional in tunnel valleys	78..
Eroded or covered after the Saalian	
till plateau/drumlin/flute eroded	8001 – 8099
push moraine/sandur eroded	81..
till plateau/drumlin/flute covered	82..
push moraine/sandur covered	83..
push moraine/sandur covered of overridden push moraines	84..
Other	
mosozioc/paleozoic outcrops topogr obstacles	9001 – 9099

Table 2.1
Labelling of the glaciogenic features in the GIS.

2.4.1 Labelling the polygons as glacial features

First, all the features were digitized in one single layer using ESRI Arcinfo Workstation. The main sources for this were previous maps of glacial features in the Netherlands (e.g. Van den Berg & Beets, 1987), geological maps (the Netherlands: Ter Wee et al 1987; Van den Berg & Den Otter, 1993; De Mulder et al., 2003; Busschers et al., 2008; North Sea: Joon et al,. 1990; Laban, 1995; Germany: Geologische Karte der Bundesrepubliek Deutschland 1:1.000.000; LBEG, 2006; Winsemann, 2009; 2010), the STRM and AHN datasets and the DGM.

In the attribute 'GLAC_ID' the type of glacial feature was labelled for each polygons. The main classes of features can be seen in table 2.1. The label consists of four digits; the first one represents the main category, the second number represents the type of feature within this category and the last two numbers are an id-number. In some cases these id-numbers were assigned to individual features, and in other cases to groups of features that are considered as one entity. For example, all the till patches of a larger dissected till plateau have the same feature label.

The main categories were split up first, according to their chronology in the glaciation; intramarginal (overridden) ice-pushed ridges, extra-marginal ice-pushed ridges and readvance ice-pushed ridges could be distinguished in this way, the same was done for the adjacent glacial basins. This was done because each type of ice-pushed ridge has different implications for the phase model; an intramarginal moraine is formed early during the glaciation, whereas the readvance ice-pushed ridges were formed during several retreats at the end of the glaciation. Some other distinction was made for features without major consequence for the chronology of the glaciation model. For example the type of till plateau: consisting of till or only remnant till, Eskers: in a tunnel valley or present as a ridge on the subsoil.

Some polygons had to be attributed to more than one glacial feature, in some cases because of the superposition of two features in other cases because of younger glacial erosion of an older feature. For example, a sandur can partially form on a ice-pushed ridge. These features were labelled as sandur on ice-pushed ridge. To avoid confusion, this feature is put into the main category that corresponds to the youngest feature.

A distinction was made between intramarginal and extramarginal river valleys. Intramarginal river valleys are formed during deglaciation, i.e. behind the ice front. Extramarginal valleys were formed in front of the ice sheet. Valleys can either be depositional (for example in glacial basins – labelled as depositional deglaciation river in a glacial basin) or eroding (e.g. till plateaus – labelled as eroding deglaciation river, eroding a till plateau). Overridden proglacial valley features that formed during the advance of the ice front were digitized into a separate layer, to prevent the GIS to become too complicated.

2.4.2 Adding additional information

Additionally, for some polygons the name of the feature was added and the authors that described it. In this way the GIS can serve as a more complete database. Some features are incorporated as lines instead of polygons in a separate GIS files. For example to visualize directional information from the fabrics and the flutes.

2.4.3 Labelling according to classical phase models

The features were also labelled according to the different existing phase models. This way, the classical phase models can best be compared relative to each other and to the new phase model. Displaying the features this way, it can easily be seen which features were taken into account by the authors and which were not. Invented non-existing features could be visualized (e.g. presumed ice-pushed ridges formed in an early phase and reworked later on).

For demonstrating how the labelling took place, the Van den Berg & Beets model is taken as an example here. The model consists of two phases, features that are not taken into consideration are labelled as '0'. Features that were taken into account started with the '2' (sequence number of the phase model) followed by the phase in which they were formed in this model. For example, a ice-pushed ridge formed in phase 1 was given '21'. Some features were formed as ice-pushed ridges in phase 1 and

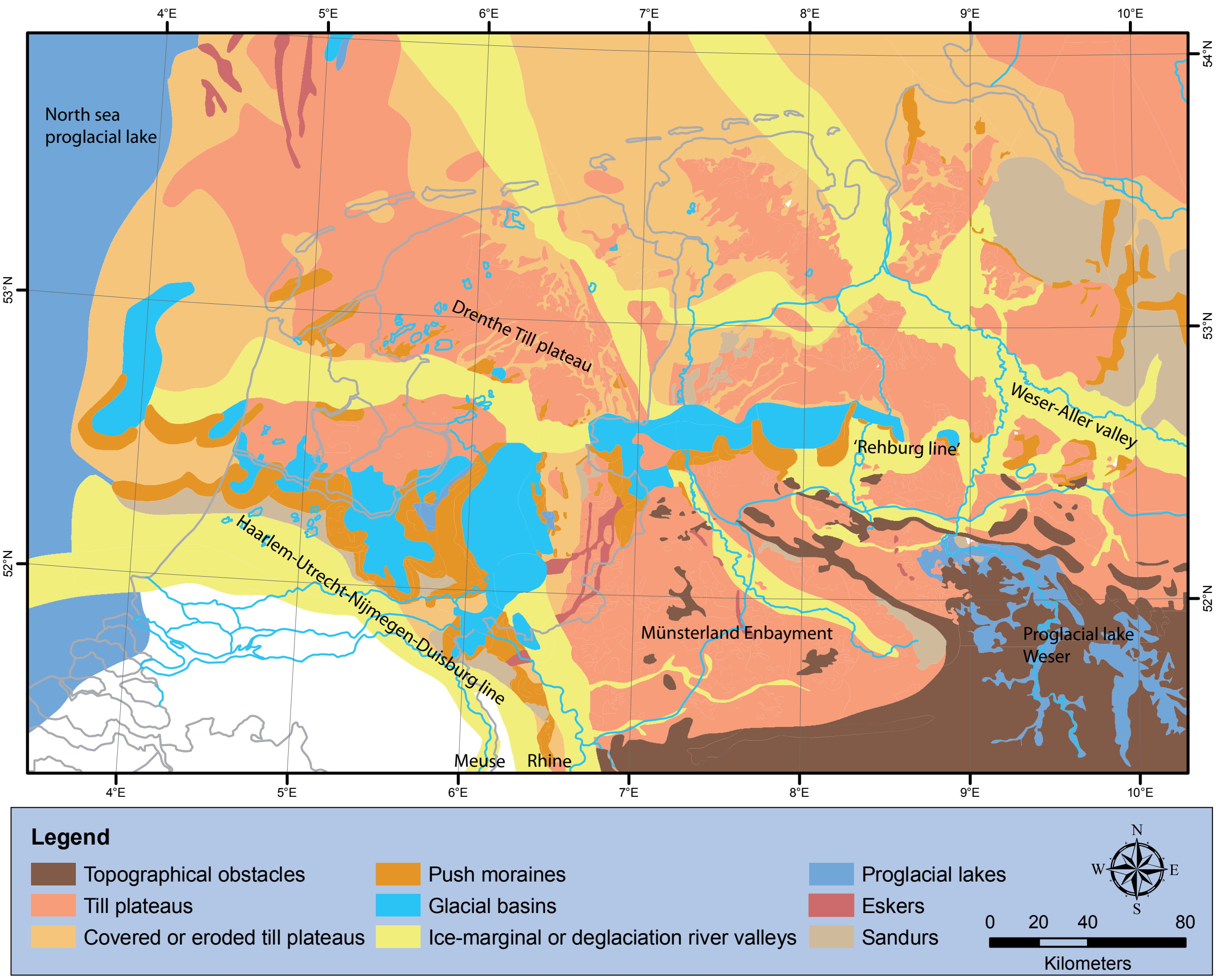


Figure 2.5
The derived simplified glacial map, some of the main topographical features are labelled, for more detailed location of features see figures in chapter 6.

overridden as drumlins in phase 2, which were labelled as '212'. Features that were present in phase two, but not actively formed (e.g. ice-pushed ridges in the central pats of the Netherlands) were only labelled as '21'. I.e. when a feature was covered by dead ice or present next to it, the feature was not considered to be formed during that dead ice phase and not labelled according to this phase.

2.4.4 Labelling according to a new phase model

A second step was to label the elements according to a new developed phase model. Features that were formed only during phase one were labelled as '1'. When a feature formed during phase 1,2 and 5 its label is '125'. As with the classical phase models, only the features that were actively being created by the glacial and proglacial processes in a certain phase were labelled as such.

2.4.5 Yielding the thematic maps and phase model maps

From the GIS queries were made to display desired thematic aspects of the mapping and phase model reconstruction. For example, all ice-pushed ridges could be displayed in one map, and their names and different types could also be visualized (figures in chapter 6). This was done for all the features to give a good overview of their presence. Queries were also made for the phase models. All the features that were taken into account in the phase model could be displayed for every single phase. This was not only done for the classical models, but also for the new model.

2.4.6 Iterative adapting of the GIS and phase models

Besides yielding these results, displaying the GIS derived thematic maps and phase models was also crucial to verify the polygons and the labels, to trace errors and to test how realistic the models were. Based on this some adoptions were made in the GIS in order to improve the new phase model and the GIS itself (figure 2.1). After this, the GIS was continuously adapted when new facts or insights arose. The GIS setup also allows these adaptions for the future.

3. Geographical and geological setting

3.1 Geographical setting

The research area covers the formerly glaciated area in the Netherlands, northwestern Germany, and part of the Dutch North Sea. The most prominent preglacial morphological feature is the Weserbergland consisting of the Teutenborgerwald and Wiehengebierge (figure 1.3). The most prominent glacial features are the large ice-pushed ridges in the central part of the Netherlands and the Rehburg line of ice-pushed ridges in Germany and their adjacent glacial basins (figure 2.5 between 52 and 53 N latitude). Besides, smaller ice-pushed ridges occur between the Elbe and Weser, in the Northern and Eastern Netherlands and the Lower Rhine Embayment (LRE). In Drenthe and large parts of Lower Saxony till plateaus with glacial lineations are common. They are often eroded by rivers that started to form during the Saalian deglaciation and continued in the Weichselian (e.g. the Weser-Aller, Elbe, Vecht, Hunze/Ems). Outwash plains (sandurs) can be found in the central part of the Netherlands, in the German Hümmling region and the Lüneburger Heide, in most cases on the distal side of ice-pushed ridges. During deglaciation large amounts of meltwater were drained trough channels that caused substantial erosion (figure 2.5). The Saalian glaciation and deglaciation were very important in forming the landscape of this region. However, before the processes before the Saalian are also relevant to the processes of the Saalian glaciation (e.g. substrate conditions). Therefore the geological development of the area before the glaciation is outlined in this chapter.

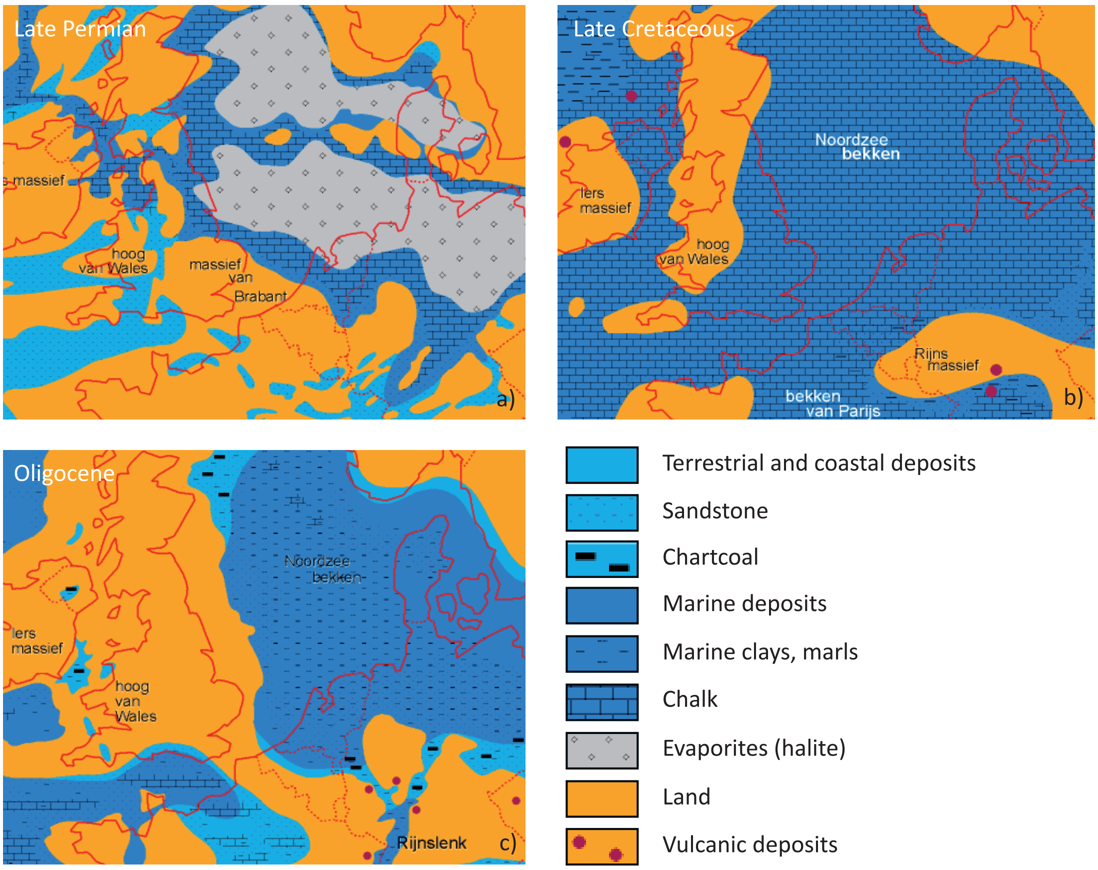


Figure 3.1
Paleaogeography in northwestern Europe. a) In the Late Permian the largest part of the research area was covered by evaporates, that later formed the salt domes. b) During the Late Cretaceous the area was covered by a shallow sea in which chalk was deposited. c) In the Oligocene, a shallow sea was present as well in the research area, except for the Münsterland Enbayment area. In the sea clayey deposits were formed. After: Ziegler (1990).

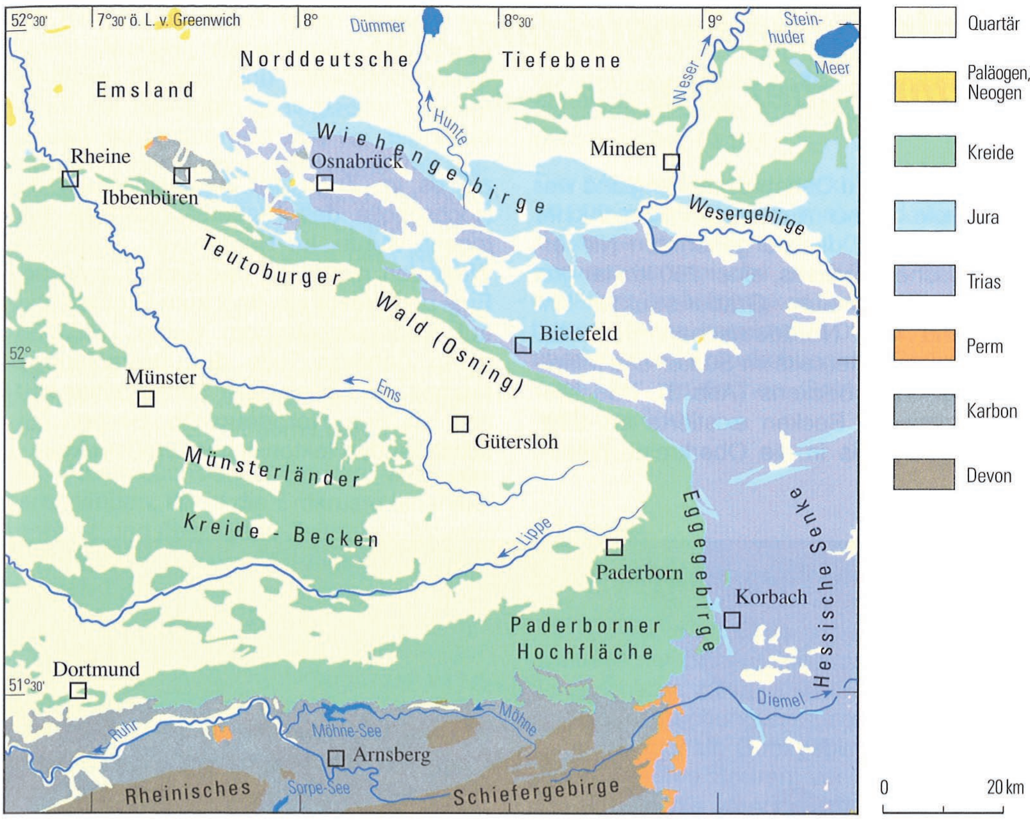


Figure 3.2
Simplified geological map of the Münsterland Enbayment and the Weserbergland. In other parts of the research area, the pre-Quaternary deposits are positioned much deeper. In the Münsterland Enbayment Cretaceous deposits dominate, Jura deposits can mainly be found in the Wiehengebirge. Also older deposits (Permian, Carboniferous and Devonian) are present there. Deposits from the Tertiary can be found on the western edge of the Münsterland Enbayment and in the Dutch eastern part of the Achterhoek and Twente. from: Groß-Dohme (2003).

3.2 Paleozoic, Mesozoic and early Cenozoic

During the Permian, the North-Sea basin started subsiding, followed by crustal extension during the Late Triassic and Early Jurassic. This crustal extension was related to the rifting that eventually led to the opening of the Atlantic Ocean and resulted in graben formation along the north-south axis of the North Sea basin (De Mulder et al., 2003). In the Late-Permian (Zechstein) large parts of NW Europe were covered by a shallow sea (figure 3.1a) in which evaporites were formed. These salt layers are present in large parts of the Northern Netherlands and NW Germany. Following their burial, the Zechstein salt has transformed into diapirs (figure 3.4) were formed after the Permian (chapter 3.3). During the Triassic sandstones and limestones were formed that nowadays crop out between the Teutoburgerwald and the Wiehengebirge (figure 3.2).
During the Jurassic, limestones and mudstones formed. These form the high ridges of the Teutoburgerwald and the Wiehengebirge today (figure 3.2). During the Cretaceous (see figure 1b) the Pompeckj Block in Northern Germany was subsiding (Szeder & Sirocko, 2005), allowing the sea to deposit mainly chalk resulting in marlstones and light chalkstones (Schreidbkreide). These deposits occur at very shallow depth at the surface in the Münsterland Embayment and form the hills of the Baumberge and Beckumerberge (figure 3.2).

3.3 Neotectonic situation

The Netherlands is located in a tectonically active region, which can be concluded from a wide variety of evidence including earthquakes, geomorphology and tilted former river valley surfaces (Van Balen et al., 2005). Tectonically, the Netherlands are located between the Mesozoic rift systems of the southeastern North Sea Basin and the Cenozoic Lower Rhine Graben (Ziegler, 1992). The main part of the Netherlands is located in the North Sea Basin. The central Dutch basin (including the Zuiderzee Basin), the Lauwerszee Trench and the Lower Saxony Basin are the main areas of subsidence in the Dutch part of the research area. The Texel-IJsselmeer High, the Groningen High and the Peel Block are stable highs (figure 3.5a). A very tectonically active region occurs around the SE-NW trending faults of the Peel-Venlo block ('Peel Horst' - Geluk et al., 1994; Cohen et al., 2002). From Haarlem to Düsseldorf this tectonic structure lines up with the maximum extension of the Saalian ice sheet (Van Balen et al., 2005). The Lower Saxony basin continued eastward into Germany (figure 3.5b). In the NW the tectonically more stable Pompeckj block is located.

In addition to the fault movements, the presence of salt diapirs since the Mesozoic affects the geomorphology in the northern and eastern Netherlands and in large parts of Lower Saxony (Baldschuhn et al., 2001; Sirocko et al., 2002). Especially in the Pompeckj Block north of Bremen some large elongated N-S oriented salt domes occur (figure 3.5b). Salt diapirs are intrusions of the salt into the younger sediment strata. The formation of these diapirs is due to the fact that salt has a low density and that it is able to behave like a plastic when a large amount of pressure is exerted. This only happens when unequal pressure is exerted onto the salt layer, then the salt tends to flow to the area with the lowest pressure. This can happen when a fault is present beneath the salt layer. The sediments on top of the salt layer will not have the same thickness, causing a difference in pressure exerted on the salt. The salt first forms an anticline, and when the pressure is high enough a diapir will form (figure 3.4). The rate at which the salt flows, is estimated to be very low: 0.01 mm/yr (Remmelts, 1992) and 0.02 (Bregman, in prep). In the Lower Saxony basin the salt domes developed during the Upper Cretaceous (Jaritz, 1973), whereas in the Pompeckj block this started already during the Triassic (Jaritz, 1992).

During the glaciation some of these old tectonic systems and halokinesis may have been enhanced

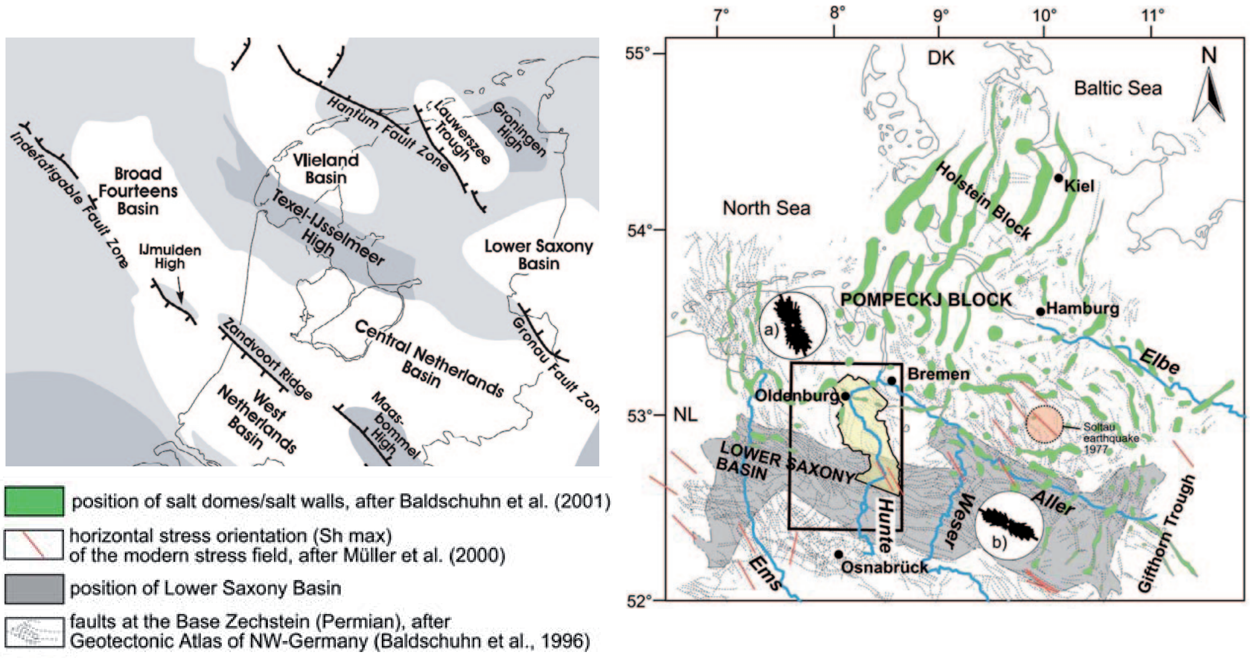


Figure 3.3 a Left: Structural outline of the Netherlands (from: Van Balen et al., 2005). White colour represents basins, shades represent structural highs. Right: structural situation of the NW-German Basin (from: Szeder & Sirocko, 2005). The position of salt domes is given in green. The faults at the Base Zechstein (Permian) are shown in black (dashed lines). The position of the Lower Saxony Basin

by the weight of the ice. These movements are not considered to be neotectonics and are called glacial tectonics (Thorson, 2000) and glacial induced tectonic activities (Bregman, in prep.). The presence of salt domes may have influenced the ice sheet by changing the hydrology, thermal regime or the substrata. This hypothesis is explored in the research of Bregman (prov. Drenthe / UU).

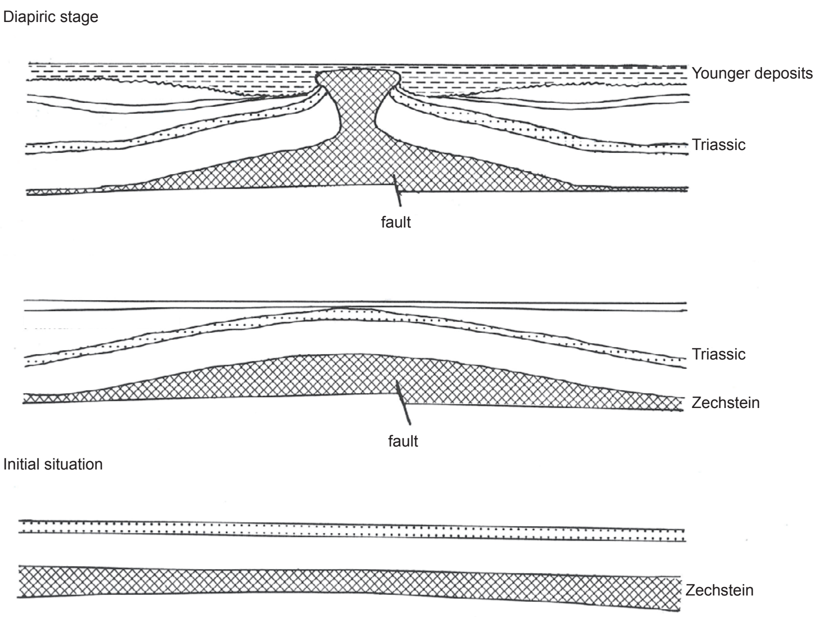


Figure 3.4
The formation of a salt dome induced by a fault (from: Remmelts, 1992).

3.4 Neogene and Early Pleistocene

In the second part of the Tertiary (Miocene), alpine mountain building had caused a tectonic reorganization which caused former Eocene and Oligocene depocentres to stop to subside and trap sediment. Consequently, terrestrial, coastal and shallow marine conditions prevailed in Germany (Skupin; 2003b; Szeder & Sirocko, 2005) and the southern Netherlands. Marine conditions prevailed in the actively subsiding parts of the North Sea Basin such as the Northern Netherlands. Several transgressions and regressions occurred within the Tertiary. This explains the alternation of shallow-marine sands and deep marine clays and the spatial and temporal variation in their distributions. The total thickness of these sediments varies from over 1000 m near Hamburg to ~400 m under Drenthe. Locally, this thickness can vary strongly due to the influence of salt domes (Szeder & Sirocko, 2005). The Tertiary fine grained deposits form a continuous hydrogeological base in the Netherlands (Van

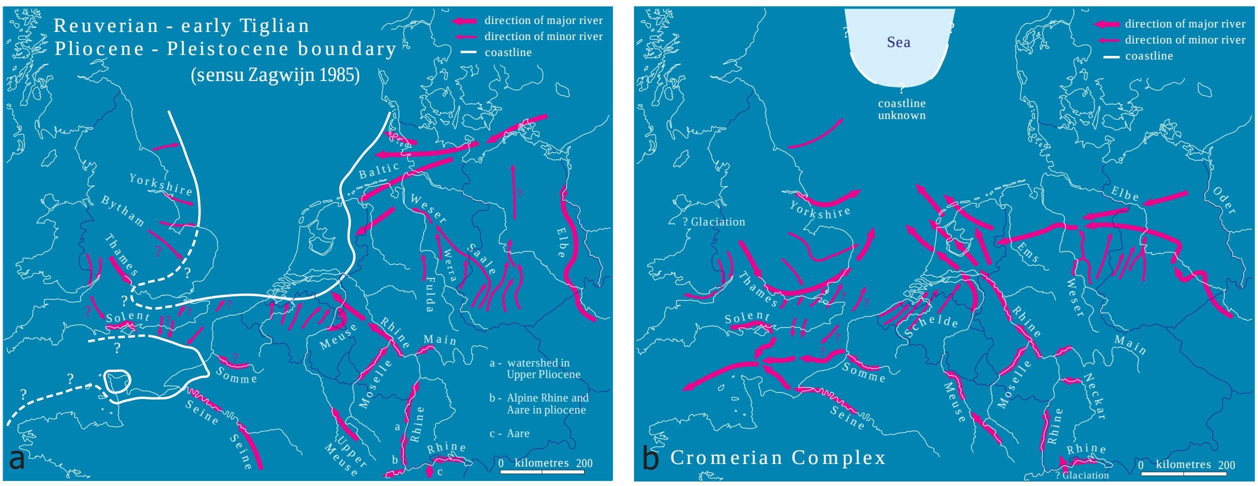


Figure 3.5
a) Palaeogeography of the river courses in northwestern Europe in the Late Pliocene and Early Pleistocene. The Baltic sea does not yet exist, the Weser and Elbe are tributaries of the Baltic river system. The Rhine already drains the Alpine region. b) Palaeogeography of the Cromerian complex (early Middle Pleistocene). The Baltic river system is replaced by the Weser and Elbe which have a much smaller catchment (After: Gibbard, 1988).

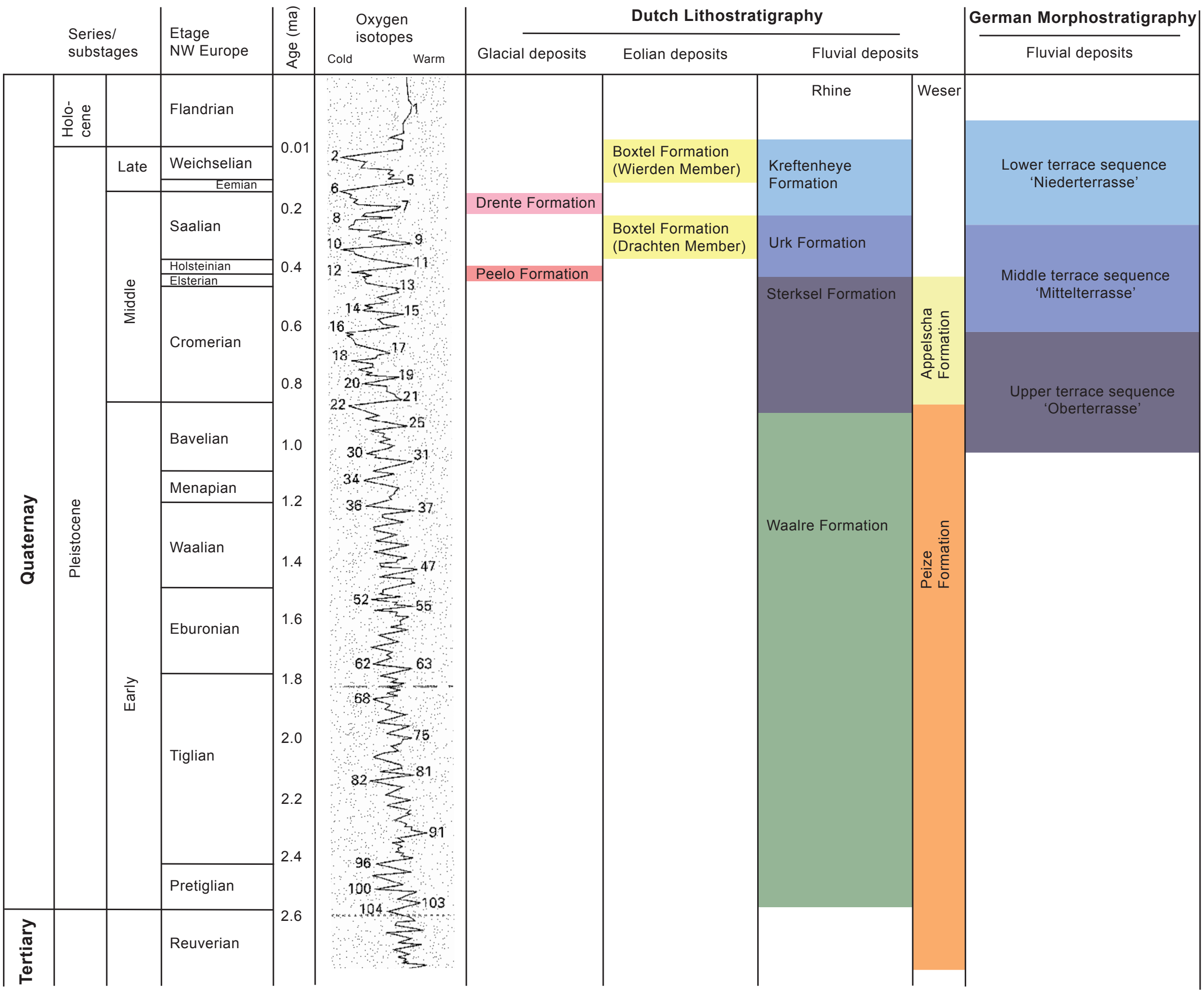


Figure 3.6
Chronology and lithostratigraphy in the research area, the formations mentioned in the text are displayed. Dutch lithostratigraphy after Weerts et al. (2000); Westerhoff et al. (2003). Germany LRE morphostratigraphy, after Boenigk & Frechen (2006).

den Berg & Beets, 1987). The Münsterland Embayment and Teutoburgerwald however became the topographical heights they are today. The Weald-Artois anticline to the southwest of the study area rose above sea level during the Tertiary, eventually disconnecting the North Sea Basin from English Channel embayment.

During the second part of the Tertiary, a large delta system fed by a large river that reached into the Baltic area build out into the North Sea basin. The Northern Netherlands and NW Germany cover the southern half of this long-lived delta system (figure 3.5a). This 'Baltic' river system (Bijlsma, 1981) is nowadays referred to as the Eridanos river system (Overeem et al., 2001). Deposits from this river system have upstream continuation into Lower Saxony and Bremen (Ehlers, 1996). In the last part of the Tertiary (Pliocene) global climate change commenced that would lead to the ice ages. The gradual closure of the Panama isthmus affected ocean currents, geared up the strength of the gulf stream in the North Atlantic and increased precipitation over Northern Europe (Haug & Tiedemann, 1998). This enlarged the discharge of the Eridanos system and the extent of deposition in the study area, it also affected the development of ice sheets (chapter 4.1).
The deposits from this river system are known as the Scheemda and Peize Formations in the Netherlands (figure 3.6 Weerts et al., 2003; Bosch, 2003a; Westerhoff et al., 2003). They account for ~75% of the

infilling of the Zuiderzee basin and consist of coarse white sands (90% quartz) and typically have a very high permeability (Van den Berg & Beets, 1987). The Balk Member was formed during Early Pleistocene high stands, and consists of beds with coastal and fluvial clays.
The progradation of the delta of the Baltic and eastern rivers continued until the Cromerian 1 ma (figure 3.5b – 3.6) (Gibbard, 1988; Westerhoff et al., 2003). Vertical aggradation kept pace with subsidence (Van den Berg & Beets, 1987). Sediments from the Rhine and the Meuse interfinger with the 'Baltic' sediments in the central and eastern Netherlands (Bijlsma, 1981; Bosch, 2003a).

3.5 The Middle Pleistocene

Since approximately 800 ka the duration of the ice ages became longer and colder (chapter 4.1), this made it possible for the ice sheets to grow larger and to extent further. This caused the Eridanos river system headlands to be demolished (Bijlsma, 1981; Gibbard, 1988). The north German basin between Baltic and North Sea became a terrain with ice-marginal river systems (Dutch: pradolina, German: Urstromtal) with much smaller catchments than the Eridanos river system (figure 3.5b). The deposits from these rivers group into the Appelscha Formation in the Netherlands (figure 3.6 - Westerhoff et al., 2003; Bosch 2003b). This formation consists of coarse and very coarse highly permeable greyish and whitish sands. In the Netherlands only small parts of this formation have remained after erosion and make up a small proportion of the later ice-pushed formations only (figure 3.12 - Van den Berg & Den Otter, 1993; Bosch, 1992; Bosch, 2003b). Probably these deposits also covered large parts of the northeastern Netherlands (Bosch, 1992). In Germany, these sands can be found north of the Wiehengebirge, around the current Rehburg line (Van der Wateren, 1995).
In the southwestern part of the research area, Rhine deposits prevail in this period. The Rhine drainage basin had grown to more-or-less its present size (connecting to the Alps) and gained discharge and sediment supply that way. The deposits form the Waalre and Sterksel Formation (Westerhoff et al., 2003; Westerhoff & Weerts, 2003). Both formations consist of coarse to fine sands with clay layers, the Sterksel Formation is coarser than the Waalre Formation.
Following the loss of the Baltic sediment supply, sedimentation in the eastern part of the Netherlands became more dominated by the Rhine. Its deposits consist of sands, silts and clays and their permeability is generally moderate to low. These deposits comprise the Urk Formation. The deposits in the Sterksel Formation equate to the Jüngere Hauptterrassen 2-4 of the Lower Rhine Embayment in adjacent Germany (Klostermann, 1992; Westerhoff, 2003; Boenigk & Frechen, 2006). The deposits of the Urk Formation are formed from ~450.000 years ago up to the Drenthe glaciation (based on ad-mixed volcanics). The Urk-Kreftenheye distinction is based on ad-mixed erratics and cross-cutting relations with of the ice-marginal features. The deposits are correlated to the gravels of the Mittelterrassen in Germany (Klostermann, 1992; Bosch et al., 2003; Boenigk & Frechen, 2006) as their upstream continuation.

3.5.1. Glaciations before the Elsterian

It highly unlikely that glaciations extended into the research area before the Elsterian. The main reason is that unequivocal glacial sediments older that the Elsterian have never been found (Ehlers et al., 1984). In northern Germany, no glaciogenic sediments are known below Elsterian strata, and they lack tills or erratics. However some pre-Elsterian fluvial deposits containing erratics have been found in the research area, indicating the proximity of an ice sheet.

The *Hattem Beds* (traditionally attributed a Mepanian age, between 1200 and 1000 ka) can be found both exposed in the Rehbuger end moraines (Van der Wateren, 1995) and in the Dutch Veluwe (Zandstra, 1971). In the Netherlands the Hattem Bed are used to define the top of the Peize Formation (Weerts et al., 2000; Bosch, 2003a). It consists of a boulder bed with gravel from the Weser catchment, but also Scandinavian erratics (Zandstra, 1971). This indicates the presence of an ice sheet near the research area (Ehlers et al., 2004; Ehlers & Gibbard, 2007) matching the story of destruction of the large Baltic river system (see above).

The *Weerdinge Beds* (traditionally attributed a Cromerian age, between 850 and 475 ka) can be found both in Germany and the northeastern Netherlands and are interpreted as distal ice-marginal glaciofluvial sediments (Ehlers et al., 1984). They contain very coarse sands from eastern provenance (Skupin et al., 1993). In the Netherlands the beds together are a member of the Appelscha Formation (Westerhoff et al., 2003; Bosch, 2003b). It is associated with a glaciation at te end of the Cromerian (Ehlers, 2005).

3.5.2. Elsterian and Holsteinian

The Elsterian (equivalent to MIS 12; 475-410 ka) is the oldest extensive glaciation in NW Europe. In the most parts of the research area, the Saalian ice reached further south than the Elsterian. On the North Sea and in Central Germany it was the other way around and the Elsterian is generally regarded the largest of the two (figure 1.3 and 3.10).

Convincing glaciogenic features of Elsterian age are deep incisions present in the substrate of the North Sea (Laban, 1995; Kluiving et al., 2003), the Northern Netherlands (Bosch, 1990) and Germany (Ehlers et al., 1984 - figure 3.7). These clearly overdeepened features reach down to 400 m below MSL in the area around Hamburg (Ehlers, 1990) and in the central parts of the North Sea (Laban, 1995). They are filled with deglaciation products and marine deposits. The fills form a complex anastomosing pattern of closed elongate depressions (Huuse & Lykke-Andersen, 2000). The channels run towards the ice margin, where they suddenly seem to stop. There is much debate about the way these depressions formed. They are thought to be subglacial channels formed by the direct flow of the ice or by its presumed basal meltwater (Kristensen et al., 2007; Praeg, 2003). The position of the multiple generations of these channels was probably influenced by the presence of the salt diapirs (Piotrowski, 1993). Relict Elsterian ice-pushed ridges were found in the north sea, east- central Germany (Thuringen) and Poland. In east central Germany the maximal extension of the Elsterian ice is beyond the extension of the Saalian ice (figure 1.3). However, in study area glaciogenic Elsterian deposits are strikingly absent, except for the bottom of the tunnel channels. This may be due to the fact that they have been eroded later on during the Saalian glaciation (Van der Wateren, 2003).

During the deglaciation lakes formed in these huge depressions, they continued to exist long after the deglaciation. The oldest layers contain dropstones mixed with sands. The largest infilling are varved silts and clays known as 'Lauenburger Ton' (Lauenburger Clay) in Germany (Kuster & Meyer, 1979). In the Netherlands this clay is called 'Potklei' (Ter Wee; 1979; Bosch, 1990) and it forms the Nieuwwolda Member within the Peelo Formation (Westerhoff et al., 2003; Ebbing, 2003). These glaciolacustrine deposits can be as thick as 150 m and in the region of Hamburg the infilling of the depression was estimated to have lasted 2000 years (Ehlers, 2005). The fine clastic fills of the tunnel valleys form a hydrological barrier

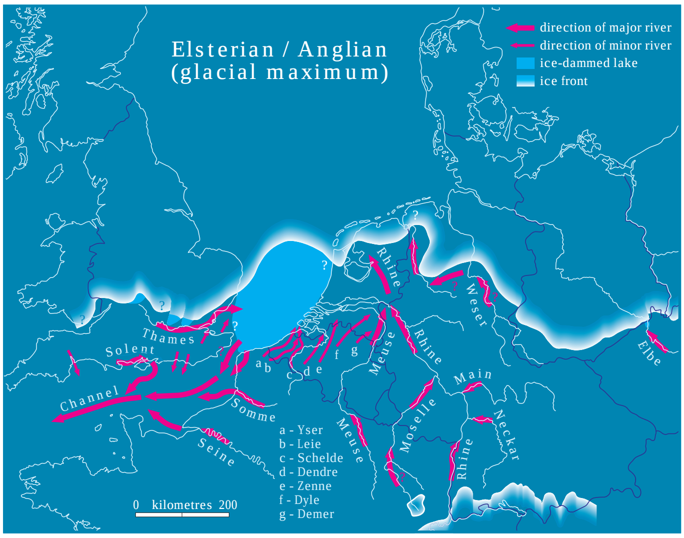


Figure 3.7
Palaeogeography of northwestern Europe during the Elsterian. The ice reached the research area but not as far as during the Saalian. In the North Sea basin probably a large proglacial lake formed (after Gibbard, 1988).

of very low permeable sediments in the pre-Saalian subsoil (Van den Berg & Beets, 1987). The upper part of these deposits consists of fine sands with glimmers (Ter Wee, 1979; Bosch, 1990). These sands may have a Saalian age instead of an Elsterian age (chapter 6.5.1).

During the Holsteinian interglacial the climate was slightly warmer than today. This period was correlated to MIS11; 410-370 ka, by Westerhoff et al. (2003) and to MIS 9 by Litt (2007). Parts of northern Germany were covered by the Holsteinian sea, many of the tunnel valley lakes became elongated marine embayments. In central and eastern parts of the Netherlands, separate river plains of the Rhine and the Meuse existed. They may have confluenced in the northwestern Netherlands or further offshore (Busschers et al., 2008). In the western part of the Netherlands these deposits can be found somewhat deeper due to the higher subsidence rate of the north sea basin in this area (Van den Berg & Beets, 1987).

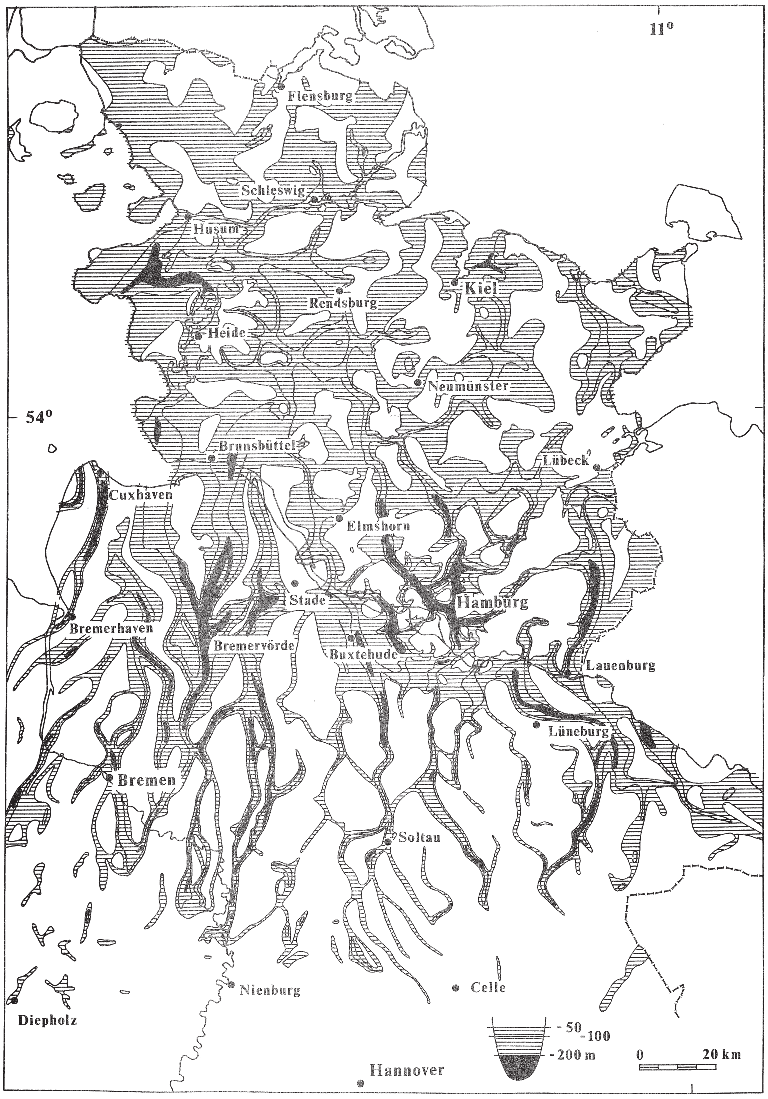


Figure 3.8
Elsterian tunnel valleys in northern Germany. From Ehlers et al., (1984).

3.6. Saalian

3.6.1. The Saalian, prior to the Drenthe glaciation

During the Saalian a sequence of interstadials an stadials occured. The correlation of these interstadials with the marine oxygen isotope stages is still problematic (figure 3.10).

The oldest part of the Saalian is known as the 'Fuhne Kaltzeit' in Germany (Ehlers., 1996). This stadial was correlated to MIS 10 (Ehlers, 1996) and to MIS 8 by Litt, (2007). This cold period was followed by a warmer interval. In the Netherlands this period is referred to as the Hoogeveen interstadial (Zagwijn, 1974; Ehlers, 1996) and in Germany as the Wacken Warmzeit (also: Dömnitz Warmzeit) interstadial (Ehlers et al., 2004). It was correlated to MIS9 (Zagwijn, 1974; Ehlers, 1996) or MIS7e (Litt, 2007). Humic intercallations defining the Hoogeveen and Bantega interstadials are found in the Netherlands (Zagwijn, 1974). In between these interstadials colder environmental conditions occurred that is correlated to MIS 8 (Busschers et al., 2008).

In the Netherlands aeolian periglacial deposits were formed, which form the Drachten Member within the Boxtel Formation (figure 3.6 Schokker et al., 2003). Recently, presumed evidence for the MIS 8 glaciation extent was found 70 km north of Ameland (Beets et al., 2005). At some time after MIS 11 and before the MIS 6 glaciation the Rhine established a second valley in the central Netherlands further west than the MIS-11-valley through the north of the Netherlands (Busschers et al., 2008). Whether this course co-existed with the Rhine course through the Northern Netherlands or replaced it, is still an issue of debate. The deposits from this system are the last Rhine deposits before the onset of the Drenthe substage glaciation. They occur extensively in the ice-pushed ridges in the Central Netherlands (Busschers et al., 2008).

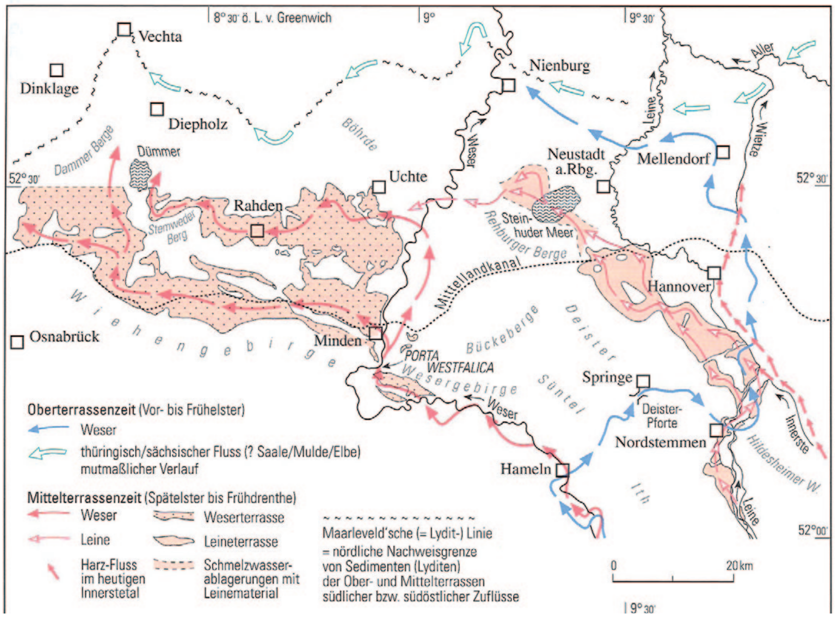


Figure 3.9
The river Weser and Leine course in the Elsterian and the Early Saalian. From: Skupin (2003a).

The Weser had its course between the Wiehengebirge and the 'Maarleveld Line' 40 km north (figure 3.9; Maarleveld, 1954; Caspers et al., 1995; Skupin, 2003a). Its deposits are called 'Oberterrasse' (Holsteinian age) and 'Mittelterrasse' (Early Saalian age), the latter contains the 'pink sands' derived from the Bundsandstein (Van der Wateren 1987; 1995; Skupin, 2003 – figure 3.8).

3.6.2 Geological and morphological situation, prior to the arrival of the ice

It is difficult to reconstruct the relief of the area before the glaciation, as it was completely reshaped during the glaciation. The few studies that address this topic, consider the topography as relatively smooth (Van den Berg & Beets, 1987), at least compared to the ice thickness (Van der Wateren, 1985). The absence of Elsterian ice-pushed structures and the wide spread occcurence of the Urk, Appelscha and Sterksel Formation support this. Also the ideas of major proglacial lakes in the Elsterian imply relative flat topography. Ongoing tectonics may have shaped the topography. As the landscape was completely reshaped during the glaciation, the relief of the area before the glaciation is difficult to reconstruct. However, it can be deduced with knowledge of the pre-Saalian deposits. The presence of salt domes may have caused some undulating relief, as the surrounding areas experience more subsidence than the domes (a hypothesis currently explored by Bregman - Prov Drenthe/UU). Residual depressions, inherited from deep Elsterian tunnel valleys could also have been present. In these depressions (proglacial) rivers had their course and probably proglacial lakes could have formed. It is however very likely, that most of these tunnel valleys had filled up to the regional groundwater base level by lacustrine, shallow marine and fluvial (figure 3.12). In Germany the Teutoburgerwald and the Wiehengebirge were topographical obstacles for the ice in the landscape. The hills marking the southern edge of the Münster Basin formed a second major obstacle, just as the topographical heights of the Baumberge and Beckumerberge - figure 1.3).

In large parts of Lower Saxony and the Northern Netherlands, thick Elsterian clayey and fine sandy deposits were present (Peelo Formation). In the Eastern Netherlands and adjacent areas north of the Variscan heights Tertiary fine sands and clays were present just below the subsurface. In the central part of the Netherlands, south of the line Almelo-Den Helder and the area around the current Rehburg line coarser grained deposits from the Eridanos rivers, Rhine and Weser (Peize, Sterksel, Urk and Appelscha Formation) were present (Van den Berg & Beets, 1987; Van der Wateren, 1995). In the Münsterland Embayment and the Westfalische Bucht mainly Cretaceous chalks were present. On the

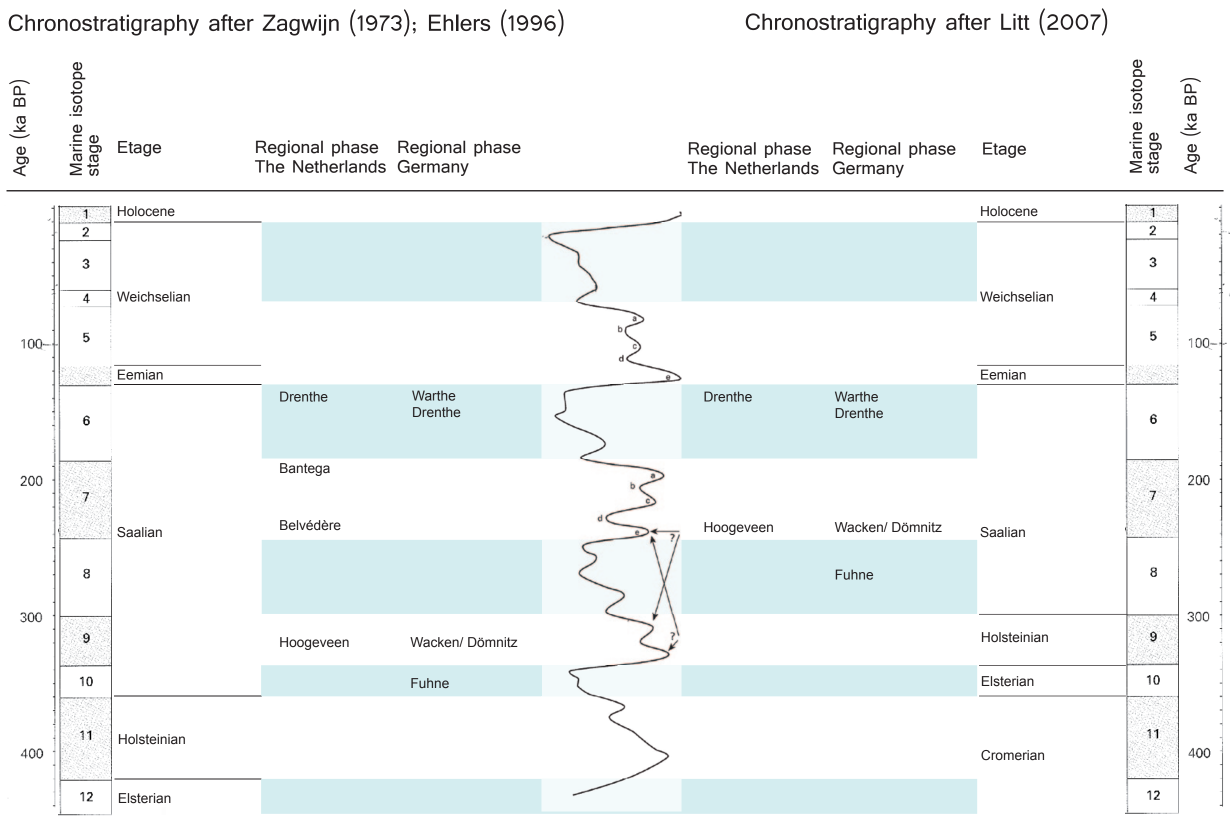


Figure 3.10
Stratigraphy of the Middle (from Elsterian) and Late Saalian. Two correlations are shown, one after Litt (2007) and Zagwijn (1974).

western edge of the Münsterbasin Tertiary deposits were present. In the LRE Rhine terraces (Upper and Middle terraces) (Klostermann, 1992; Boenigk & Frechen, 2006).

3.6.3 The Saalian, Drenthe-Warthe stage

The last part of the Saalian includes the coldest period during which the major glaciation of the Saalian took place (figure 3.10; Ehlers & Gibbard, 2004; Litt, 2007). This part of the Saalian equals to MIS 6. The glaciation is generally divided into two stages. First, an oscillatory increase in ice volume occurred from ca. 195 ka up to the Drente advance at ca. 155 ka (Lambeck et al, 2006) in which the ice managed to cover the major part of the study area (figure 3.11a). The Drente maximum stage had a duration of

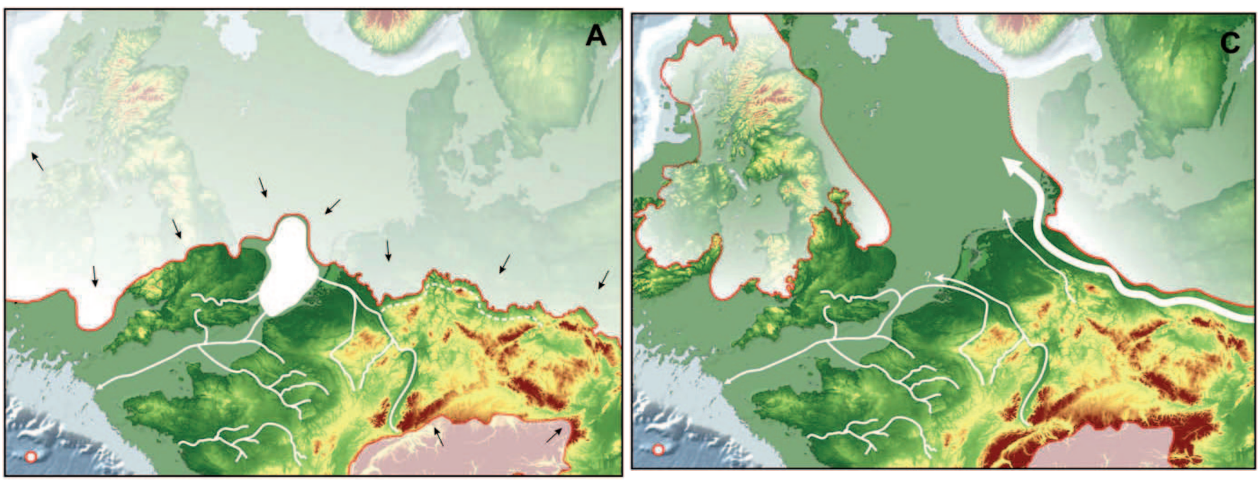


Figure 3.11
Presumed palaeography of northwestern Europe in the Drenthe (A- maximal ice extension) and Warthe substage (C). From: Toucanne et al. (2009).

ca. 5 kyr. After this phase some ice retreat occurred, which is consistent with the global sea-level rise inferred at ca. 150 kyr BP (Lambeck et al., 2006). The second stage is the Warthe stage at ca. 143 kyr BP, in which a major readvance of the ice front occurred (Ehlers et al., 2004). The Warthe maximum lasted until 140 kyr BP and was followed by rapid melting (figure 3.11b - Lambeck et al., 2006). These dates roughly coincide with those found by Toucanne et al. (2009): ~160 ka and ca 150–140 ka for the Drente and the Warthe respectively, the major deglaciation of the Drente stage happened around 155 ka – explaining an observed increase in sediment flux off the mouth of the English Channel.

3.7 After the Saalian

After the Saalian glaciation a strongly glaciation-affected landscape was left behind with considerable variations in relief. Ice-pushed ridges with crests over 100 meter asl alternated with erosive depressions with water-bodies locally over 50 m deep. During the Eemian (equivalent of MIS 5e; figure 3.10) the sea level rose to approximately the same level as today. Basins were filled in with shallow-marine deposits in the Gelderse Vallei and the areas to the northwest (e.g. De Gans et al., 2000; Ehlers, 1996; Beets et al., 2006). The Rhine valley of that time is recorded as part of the Lower Terraces in the LRE (Niederterrassen – e.g. Klostermann, 1992; Boenigk & Frechen, 2006) and occupied the IJssel tongue basin in the central Netherlands (e.g. Busschers et al., 2007). The Meuse had a separate course south of the line of the Saalian maximal ice extension.

In the Weichselian glaciation (MIS 4-2; figure 3.10) the Scandinavian ice sheet reached the very northeastern part of the study area (e.g. Ehlers, 1996; Ehlers et al., 2004; Svendsen et al., 2004). This caused periglacial conditions to prevail in the rest of the study area, promoting large braided river systems that were able to form wide terraces. The Saalian ice marginal and deglaciation river systems

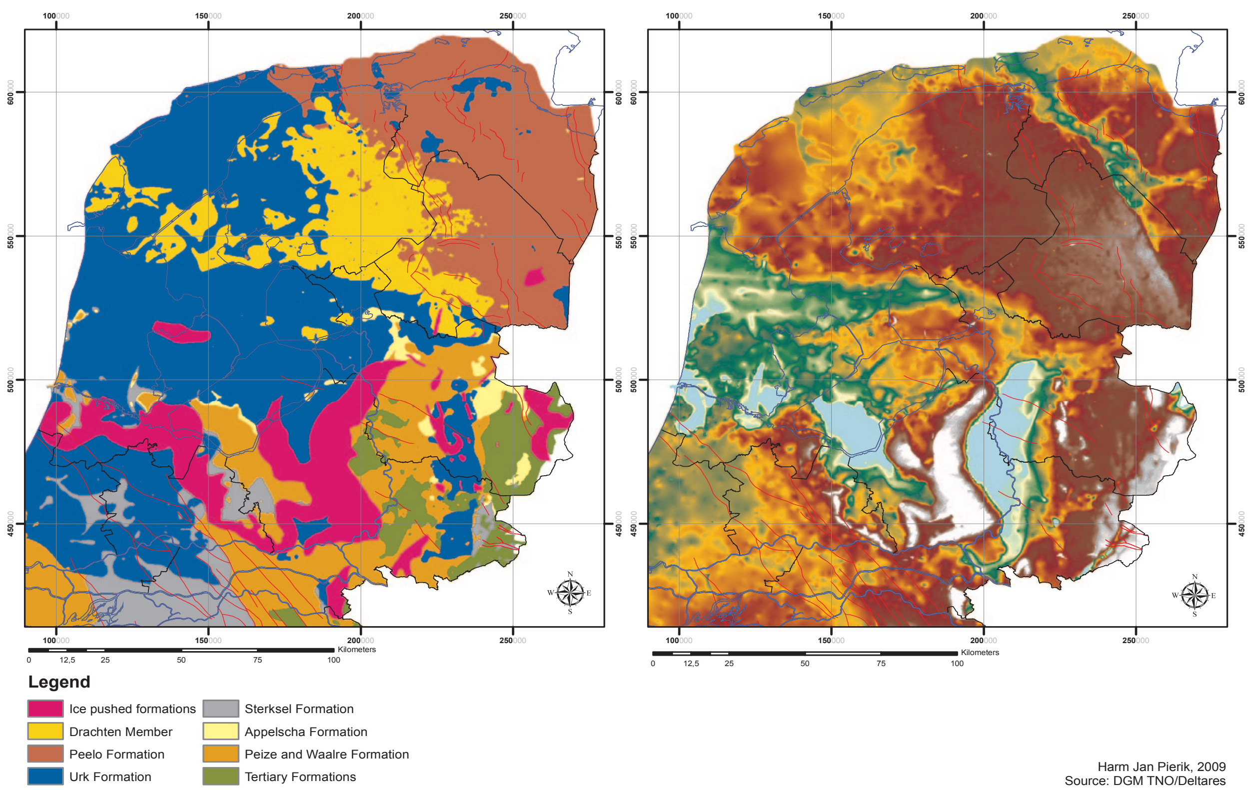


Figure 3.12
Extension of the deposits older than MIS6 in the Netherlands, the ice pushed deposits are also displayed. In the northern Netherlands the relatively fine grained Peelo Formation and Drachten Member (Boxtel Formation) are found under the tills, and are assumed to have formed the preglacial surface. In the Central Netherlands coarser grained deposits of the Peize and Urk Formation were present. In the eastern part fine grained Tertiary deposits are present, that continue in Germany. As the figure shows the current position of the pre-MIS6 deposits, erosion during the glaciation (glacial basins) and afterwards can also be seen (more info see figure 6.31).

were reoccupied by glacial meltwater and in the Netherlands relative wide terraces were formed by the Rhine and Meuse (Busschers et al., 2007). Ice-pushed ridges in the Dutch-German border region were eroded when the Rhine valley avulsed and reconnected with the Meuse tributary in the west-central part of the Netherlands from ca. 60.000 years onwards (Busschers et al., 2007). Other ice-pushed ridges were severally affected by the periglacial conditions due to solifluction, snow melt permafrost build-up and decay yielding the so called 'dry valleys' (Maarleveld, 1953; Van der Wateren, 1995). Throughout the Weichselian large amounts of areas were covered with cover sand or loess that could be transported due to the spareness or absence of vegetation (Koster, 2005). The largest Weichselian loess plateau of the research area can be found on the Syke-Cloppenburg till plateau (Skupin, 2003a). In the Holocene, the sea level rose again causing marine erosion of glacial features in the coastal areas, whereas other features were covered by marine deposits.

4. Ice sheet dynamics

4.1 The forcing of glacials

The onset of major glaciations in the Northern Hemisphere about 2.7 million years ago the result of climate cooling that set in during the late Pliocene epoch (Lawrence et al., 2006; Ehlers & Gibbard, 2007). This longer term cooling trend was induced by the closure of Panama isthmus in several stages between 3.6 and 2.6 Ma ago (Haug & Tiedemann, 1998; Bartoli et al., 2005). The closure caused an intensification of the Gulf stream, transporting more moisture to the northern latitudes. This moisture supply was a precondition for the build up of ice sheets on North America, Greenland and NW Europe. Subsequently, the Pleistocene saw a long series of repeated glacial-interglacial climate cycles (figure 3.6), during which the Northern Hemisphere ice sheets successively expanded and retreated. During at least two of them, the Scandinavian ice cap stretched across the research area. The timing of ice cap formation is generally explained by variations in the Earths orbital parameters (Hays et al., 1976). Due to configuration of the orbital parameters, the Earth periodically received less insolation during the summer, allowing ice caps to be built up.

Roughly until 1.4 Ma glacial cycles of 41 ky length occur, since 0.6 Ma these cycles are 80-120 ky, from 1.4 – 0.6 Ma a transition interval occurred. The transitions happened without an apparent change in insolation forcing (Clarck et al., 2006). Several hypotheses have been proposed to explain the transition, both including and excluding ice-sheet dynamics. Recently, Bintanja & Van der Wal (2008) concluded that this transition can be attributed to the increased ability of the merged North American ice sheets to survive insolation maxima and reach continental-scale size.

4.2 Initiation of ice sheets and its feedbacks

Where the ice sheets originated and how individual ice sheets could grow so large is explained by two theories (Ehlers, 1996). The first theory states that the glaciers started to form in highland areas. These glacier merged in the foreland and finally became a large ice sheet. They grew larger mainly in the direction to wind came from. The wind delivered the moisture required to build up the ice sheet (Flint, 1971). This theory was followed by Boulton et al. (2001) who stated that the location of nucleation centres of the Scandinavian ice sheet was determined by topography during the Weichselian. Ehlers (1996) proposed the theory of instantaneous glacierisation, that especially holds for the Laurentide ice sheet. They proposed that a large amount of precipitation of snow in a large area could also cause an ice cap to initiate.

The formation of these ice sheets caused an additional cooling (positive feedback) due to the increased albedo of areas covered by snow and ice. More solar radiation is reflected, causing the Earth to cool down even more, allowing the ice sheets to build up even more closing the positive feedback loop. This positive feedback is in function until the summer solar radiation received by the extended ice cap surface is large enough to melt the amount of annual (winter) ice sheet growth. If the amount of solar radiation increases the decay of an ice sheet can set on very quickly. Decay is faster than buildup of ice, which takes thousands, or even ten thousands of years.

Regardless of the fact that it takes time to build up an ice cap, ice fronts can prograde quite fast, especiallt when the front is lobate (meters/year). In the Saalian, the ice sheet reached the research area, located on the very southern margin of the ice sheet, only in the last period of the glaciation, after which relatively rapid deglaciation took place.

4.3 Mass balance and ice flow

4.3.1 Mass balance

When the input of snow (accumulation) exceeds the melting of the body (ablation), the ice sheet can be sustained. Ablation usually takes place due to the direct melting of the ice sheet, but iceberg calving in proglacial lakes or the sea can also occur. The melting of the ice sheet can occur at the surface, or at the base due to geothermal heat or friction as a consequence of ice flow. Most of the ablation takes place at the margin of the ice sheet. It can also occur further away from the margin, especially during deglaciation.

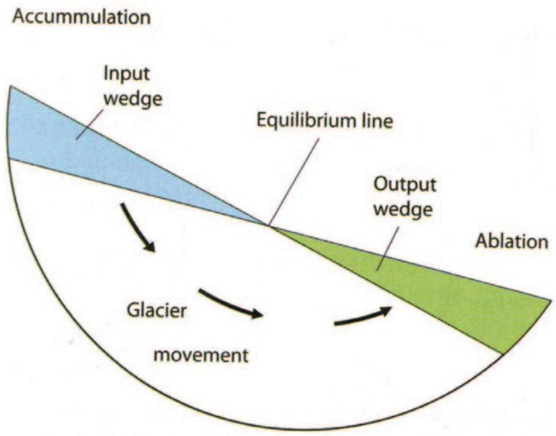


Figure 4.1
Idealised glacier showing the mass balance in an ice sheet. From Bennett & Glasser (2009); after Sudgen & John (1988).

The part with net accumulation and the part with net ablation are separated by the equilibrium line, also called the snow line. The part with net accumulation is usually found in areas either high latitude or high altitude, i.e. towards the centre of an ice sheet. The part with net ablation is located in areas with a lower latitude and altitude, i.e towards its margins. Due to net accumulation, the area above the equilibrium line tends to be elevated higher. Svendsen et al., (2004) estimated that during the LGM the ice at the centre of the Scandiavian ice sheet (Botnia Bay) was 2700 m thick, the centre of the British ice sheet was estimated to be appox. 300 m thick. On the other hand, the area with net ablation has a lower elevation. This gradient of elevation is the driving mechanism for the ice to flow. In a longitudinal section an ice sheet has a convex profile; the slope of the glacier surface is flat in the centre and steep towards the margins.

In warmer maritime climates with a lot of precipitation available (at the ice margins), accumulation and ablation rates will be high, causing the glacier to flow fast. Whereas in regions with colder and drier climates, such as the ice sheet interior, the glacier will flow much slower (Bennett & Glasser, 2009).

4.3.2 Stress and ice flow

The stress that is exerted on the ice as a consequence of flow can be expressed in the formula:

$$\tau = \varrho g h \sin\alpha$$

In which τ is the shear stress, ϱ the density of the ice, h the thickness of the ice and α the slope of the glacier surface. From this formula it can be seen that a glacier experiences more stress when the ice is thicker and the slope of the glacier surface is higher. This basal shear stress can be exerted in three ways: by internal deformation, basal sliding and deformation of the subsoil (Bennett & Glasser, 2009)

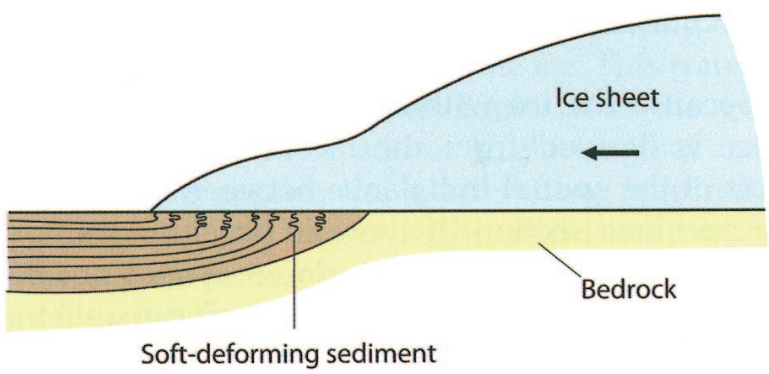


Figure 4.2
Schematic cross-section through an ice sheet illustrating the deformable bed. From: Bennett & Glasser (2009).

The internal deformation can take

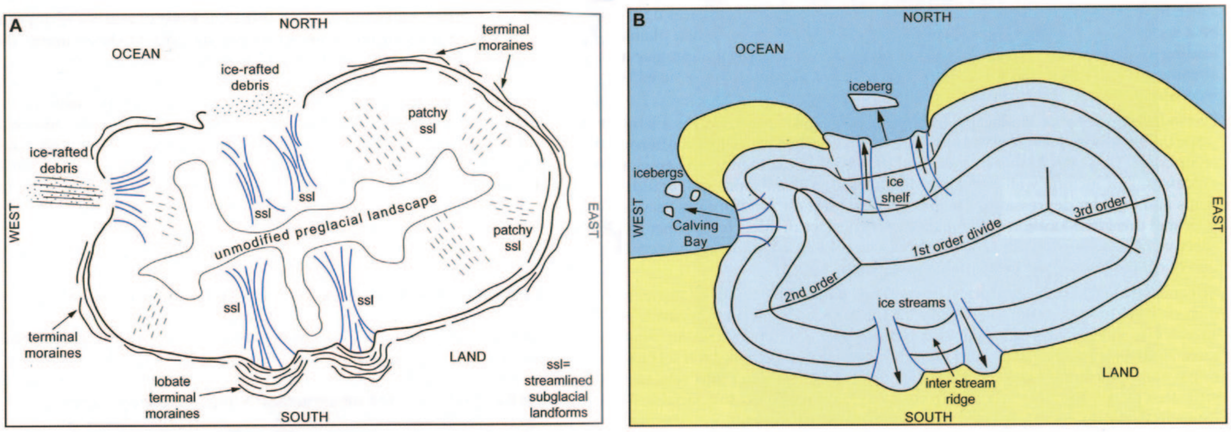


Figure 4.3
A) A schematic example of a palaeoglaciological reconstruction. B) Reconstruction of the former ice sheet based on the landform record. It shows first-, second- and third-order flow divides, ice streams, inter-stream ridges, and calving bays. The bottom margin of the ice sheet is comparable to the situation in the research area including a lobed ice margin with some distinct ice flows. The locations of divides may be controlled by the distribution of accumulation or because of drawdown by flanking ice streams, calving bays and ice shelves. From: Bennett & Glasser (2009), after Boulton et al. (2001).

place by changes in the structure of the ice (creep), folding and faulting of the underlying sediments. Basal sliding of the ice involves enhanced basal creep and regelation slip. Deformation of the subsoil takes place when the sediments beneath a glacier are unfrozen and therefore able to transport water. The pore pressure in the sediments will be increased due to the melt water which makes the sediments weaker. Under these conditions, ice-pushed ridges may be able to form (chapter 5.3 - Van der Wateren, 1995; Bakker, 2006). When the bed is deformed, this can cause the glacier to flow faster under the same shear stress conditions causing the ice thickness to decrease (Bennett & Glasser, 2009), this can be seen in figure 4.2.

4.3.3 Changes in ice flow patterns

Changes in ice flow patterns that occur on a local scale, can originate from flow changes at the scale of the continental ice sheet. This is usually associated with major shifts in the ice divide, which is defined as the line of flow divergence (Boulton et al., 2001). Its location is a consequence of patterns of accumulation on the ice sheet surface and patterns of ice drawdown in areas of fast flow. In a large

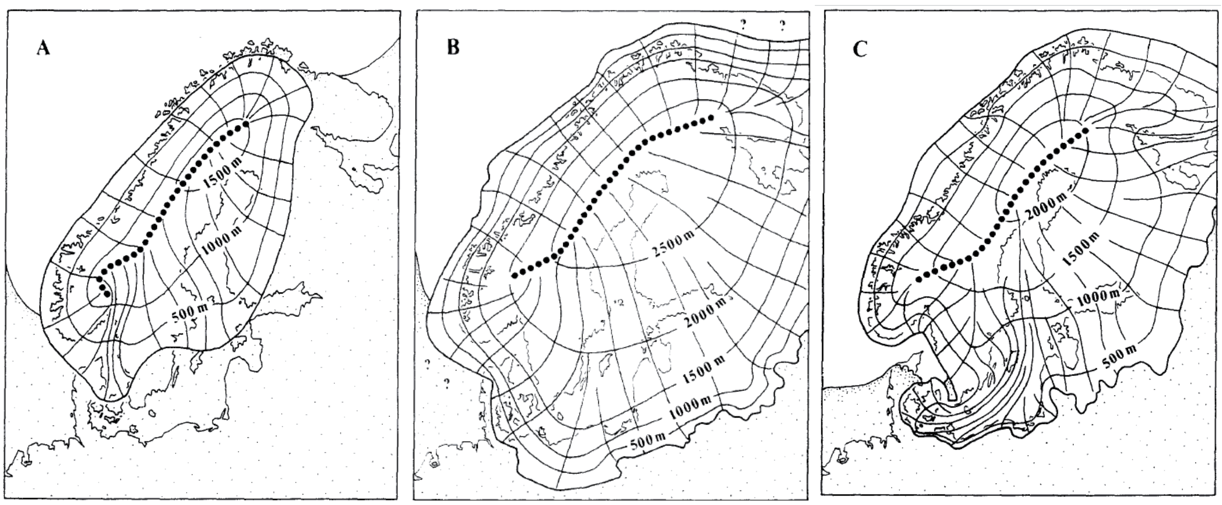


Figure 4.4
Change of the ice divide and flow patterns of the Weichselian ice sheet. A = built-up of the ice sheet, B= Weichselian maximum, C= Pommerian stage. This process causes the ice flow in Germany to deflect from the NE to the E. From: Ehlers (1990a).

ice sheet, usually several ice divides are present (figure 4.3).
In the Weichselian, the Scandinavian ice divide shifted more to the east and to the south towards the end of the glaciation. According to Boulton et al. (2001) the shift to the east is the result of warm and moist conditions on the western flank and cold and dry conditions on the eastern flank. Climatic warming leads to the decay of the western part of the ice sheet while the the eastern part may continue to grow. This generates asymmetric mass balance profiles in the direction of the wind (Van den Berg et al., 2008). Another mechanism that plays a role is the freezing of the Northern Atlantic. When the ice sheets developed, the northern parts of the Atlantic ocean froze. This caused the northern part of the ice sheet to become dryer and the ice divide to shift southward. This occurred during the Weichselian; at the beginning of the glaciation most accumulation occurred in the northeastern part of the ice sheet, while during the LGM the southwestern part grew fastest (Svendsen et al., 2004; Siegert & Dowdeswell, 2004). These two effects cause a clockwise shift in ice divide position, in turn causing changes in flow direction at the ice margin (Ehlers, 1990; 1996).

4.4 Subglacial hydrology

4.4.1 Cold and warm based glaciers

Warm-based ice is defined as being at the pressure melting point, therefore having water available at the bed, whereas cold-based ice is below the pressure melting point, with no water available at the bed (Paterson, 1994). These conditions can vary between glaciers, but can also vary temporally and spatially within one glacier and they are very important for the mechanical processes at the glacier base and shape the morphology of the subsoil (Bennett & Glasser, 2009).
The thermal regime under a glacier depends on the balance between the different energy sources and sinks and the pressure under the ice. Ice surface temperature, geothermal heat and frictional heat are the main sources of heat (Bennett & Glasser, 2009). Conditions that favour warm based glaciers are: high accumulation rates of relatively warm snow at the margins, high summer melt and a thick and fast moving ice. Besides, a spatial relation between elevated geothermal heat and the presence of salt domes appears to exist in the study area (Bregman, in prep).

The amount of meltwater under the sole of the glacier has important consequences for the glacial dynamics (e.g. ice streams and surges) and thus the morphological evolution of the glaciation area. Cold-based ice can transport subglacial sediment and erode its bed (Cuffey et al., 2000), but the amount of deposition and erosion are much less compared to warm-based ice (Kleman et al., 2006). Cold-based ice is therefore associated with preservation of older glacial landforms during younger glacial events (Waller et al., 2009). In hardrock regions, i.e. the centre of the Fennoscandinavian ice sheet, cold-based glaciers can cause erosion by plucking (Boulton, 1972). Warm-based ice is associated with more widespread erosion, deposition and reshaping of the ice sheet bed (Kleman & Glasser, 2007), and it thought to have affected out study area (e.g. Van den Berg & Beets, 1987).
The spatial distribution of cold and warm based conditions in ice sheets can be very complex. Based on geomorphological features in Scandinavia Boulton et al. (2001) demonstrated that the during the maximal extension of the Weichselian, the Fennoscandian ice sheet was cold based in its centre and mainly warm based at its margins (figure 4.3). Therefore, it is assumed that the majority of landforms in the research area were formed under warm-based conditions.

4.4.2 Groundwater flow

The occurrence of an ice sheet does not only influence the hydrology at the base of the ice, also deeper hydrology is affected. Groundwater flow patterns can even be radically changed by ice sheets, up to several kilometers deep and 40 km in front of the ice front (Boulton et al., 1995; Piotrowski, 2006). The groundwater is forced towards the ice margin, sometimes high flow velocities can occur. When the drainage capacity of the subsoil is insufficient, eskers or tunnel valleys may be produced (Boulton et al., 1995; 2009). There are also indications that this process influenced the grondwater flow in Drenthe (Bregman & Lüse, in prep.). It can ause the glacier to flow faster because of increased basal sliding of a water sheet. Pressurized groundwater may produce large scale glaciotectonics or folding.

4.5 Debris transport

Sediments that are transported by glaciers are usually unsorted and they can originate from local areas as well as from more distant areas. Supraglacial debris is located on top of a glacier, but it is relatively uncommon in large ice sheets (Bennett & Glasser, 2009). Englacial debris and subglacial debris are located respectively in the ice sheet and at its base. Most debris transport in a large ice sheet takes place subglacially. The clasts are usually better rounded than other debris (Boulton, 1978) it is also finer and a bimodal grainsize distribution is common of coarse grains and mineral sized grains (Bennett & Glasser, 2009). Most of this debris is finally deposited as till (see examples in chapter 5).

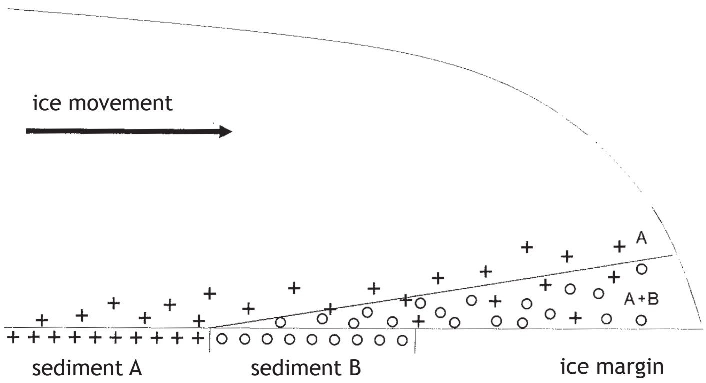


Figure 4.5
Transport in a glacier; In a till section at the margin the materials that are derived from the most distal source areas (A) are found on top. After: Rappol (1993a)

According to Rappol (1991; 1993) debris that has been transported over the longest distance is usually deposited on top of a till sequence in the marginal area. This is based on the assumption that subglacial debris is distributed gradually upwards in the ice up to some extent. The lithology that occurred near the centre of the ice sheet has been spread out vertically already at the time that the ice encounters a new lithology (figure 4.5). This mechanism to some extent can explain the till stratigraphy in the research area (see chapter 6.1).

4.6 Marginal ice streams and surging

At the margins of ice sheets, ice streams and surges can occur (figure 4.3). In the literature there is some confusion on the terminology of ice streams and surging (Clark & Stokes, 2003). Ice streams can be defined as 'a region in a grounded ice sheet in which the ice flows much faster than in the regions on either side' (Paterson, 1994) or as 'narrow fast-flowing areas within an ice sheet through which most of the ice, meltwater and sediment is discharged' (Winsborrow et al., 2010a). The term 'surge' is used to describe flow acceleration of a temporary (and non-steady state) nature. For example, a dramatic flow acceleration and possible

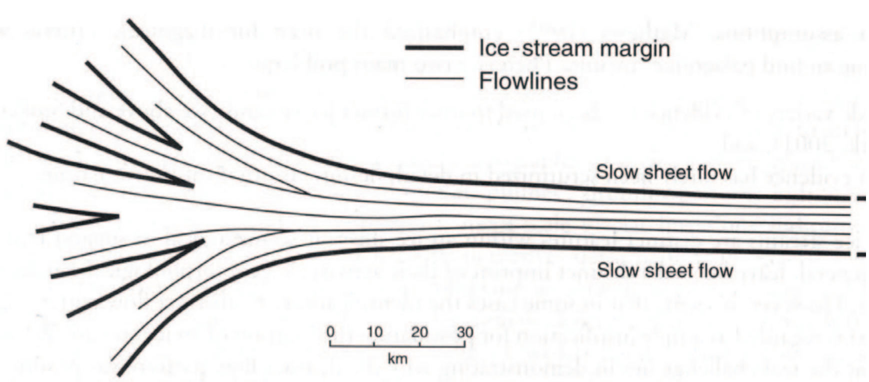


Figure 4.6 Idealized flow geometry of an ice stream. A highly convergent onset zone feeds the faster-moving ice stream trunk (Clarck & Stokes, 2003).

margin advance of the whole ice mass, and which happens at regular cycles (Clark & Stokes, 2003)

Proglacial water masses, such as ice dammed lakes and seas are considered as a major control for ice streams and surges, as buoyancy reduces the bed friction thereby enhancing ice flow (figure 4.3 and 4.6 - Boulton et al., 2001; Winsborrow et al., 2010b). Subglacial hydrology can be influenced by a geothermal source underneath an ice sheet. Increased amounts of meltwater will decrease the ice-bed friction, possibly triggering or leading an ice stream (Winsborrow et al., 2010b). Such processes are hypothesised to have operated between the salt domes and the Hondsrug ice stream (Bregman, in prep). Both surges and ice streams may show a cyclic behaviour, when the flux in an ice stream exceeds the capacity of its catchment to nourish it, the streaming ceases (Boulton et al., 2001).

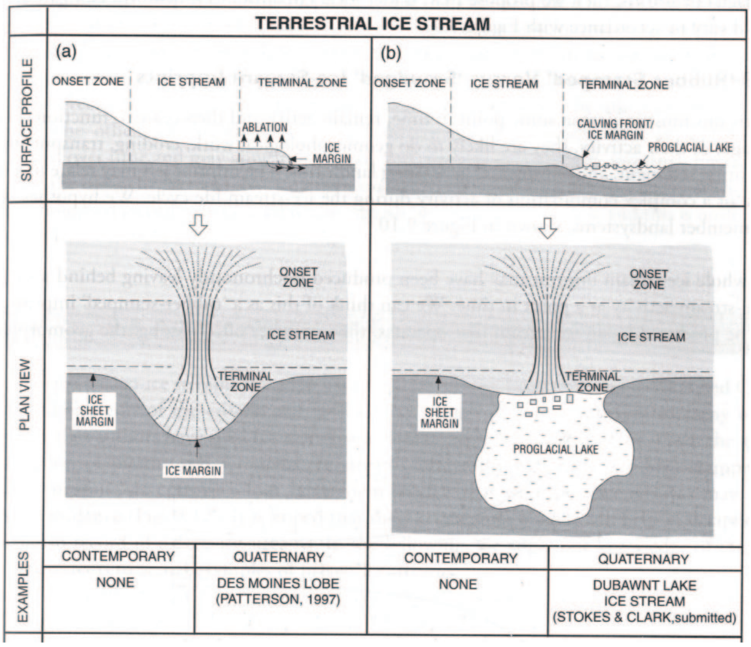


Figure 4.7
Terrestrial ice stream, ending as a lake environment resp. ending in a lake, with calved icebergs (Clark & Stokes, 2003).

Ice streams have large dimensions, up to tens of kilometers wide and hundreds kilometers long in large ice sheets (Boulton et al., 2001). They have highly convergent onset zones feeding the main channel (figure 4.6). Flow velocities may be up to 300 m/yr. They have very abrupt lateral shear margins and a spatially focused sediment delivery (Clark & Stokes, 2003). Ice streams are very relevant for the mass balance of an ice sheet, because large amounts of ice can drain an entire ice sheet through relatively small sized ice streams (Clark & Stokes, 2003, Winsborrow et al., 2010a). Therefore, these authors regard ice streams to be of fundamental importance for reconstructing ice sheet dynamics. These reconstructions can be made based on numerical modelling and geomorphological analysis (e.g. Boulton et al., 2001). The orientation and intersection relations of glacial lineations and drumlins can is used to reconstruct ice streams in the research area (chapter 2.2.1, 5.2.3).

Surges are more common in valley glaciers than in continental glaciers as the subsurface slope is much smaller for continental glaciers (Ehlers, 1990a). During a surge rapid glacier advance takes place, much faster (up to 1000 times) than under normal conditions (Clarke et al., 1984). A surge occurs as long as the meltwater supply due to thermal melting or a large pressure, is larger than the drainage capacity. The meltwater can lift the ice and a significant reduction in basal friction takes place allowing the glacier to surge. Besides, Truffer et al. (2000) argued that large scale mobilization of sediment layers beneath the ice could be a key factor in initiating glacier surges. Surges are induced by a more positive mass balance combined with available thresholds (Sugden & John, 1988). Reorganization of the subglacial drainage system can also be an important factor for inducing surging (Clarke et al., 1984).

5. Glacial, proglacial and deglacial deposits and landforms

This chapter provides background information on a selection of glacial deposits and landforms found in the studied region and comparable areas elsewhere in the world. Specific deposits and landforms in the research area are described in chapter six.

5.1 Tills

5.1.1 Definition, composition and distribution of tills

Till (Dutch: keileem; German: Geschiebelehm) is a loosely defined genetic term for debris deposits that has been directly deposited by the ice. A till is a diamicton, which is a general lithological term for a non or very poorly sorted sediment lacking internal sedimentary structure. Deposits that are created by running meltwater underneath and in front of the ice are not to be regarded as tills. Pre-glacial sediments that are structurally deformed and displaced by ice, but otherwise have their original sedimentary structure, such as in ice-pushed ridges, neither are called till (Ehlers, 1996).

A till is commonly composed of large pebbles in a matrix of clay and silt. Usually, tills are matrix supported, but the lithological characteristics and pebble density are very variable (figure 5.1). These characteristics are co-determined by the origin of a till. The pebbles and gravel in the till (chapter 4.5) can indicate sediment source areas far away from the locality of deposition (chapter 6.1). But at the same time, the till can contain large amounts of relatively local materials.

The thickest tills under an ice sheet are deposited at the margins, because transport of materials in an ice sheet is towards the margin (Boulton, 2006). Tills are present in a large variety of landforms. They occur as till sheets, e.g. the Drenthe Plateau and the German Geeste. Within a till sheet secondary landforms may be present; drumlins, flutes and rogens. Tills can also occur on top of ice-pushed ridges and in ice-pushed ridges.

Figure 5.1
Section in a till; the lithological composition is very variable.

5.1.2 Genetical till classification

Traditionally, three common genetically different types of till are distinguished: lodgement till, melt-out till and gravity flow till. Additionally, deformation tills and sublimation tills are distinguished, the latter is quite rare. Recently, this classification has been challenged by several authors (e.g. Evans et al., 2006; Bennett & Glasser, 2009), arguing that most till accumulation are formed by a complex interaction of different processes that vary temporally and spatially. Therefore, it can be impossible to distinguish the types of tills in the field or at micromorphological scale. Lodgement till, for example, is very hard to distinguish from a subglacial meltout till. They distinguish only two basal types of till, subglacial and supraglacial tills, which can be formed due to different processes (lodgement, melt-out). These till types and the processes involved are presented below. In the research area, till classification was applied mainly based on source area. That classification is outlined in chaper 6.1, this chapter covers genetical classification.

Subglacial tills

Subglacial tills comprise lodgement till, subglacial meltout till and deformation till. Lodgement till is

being deposited at the glacier sole by actively moving ice. It can be formed when the drag of a glacier is larger than the frictional force of the subsoil or when a subglacial moraine melts out due to pressure melting (Boulton 1975; Sugden & John, 1988). This can occur grain by grain. Also, sheets of till can be deposited by subsequent melt-out.
The rate at which lodgement till is being deposited depends on the basal temperature, the amount of englacial debris and the ice flow velocity. Ice velocity also determines whether a continuous till sheet is formed or only parts of glaciated areas are covered with tills in the lee sides of obstacles.
The particle distribution in a lodgement till is usually bimodal or multimodal. Clasts commonly show round edges and may contain striae. Pebbles in lodgement tills often show imbrication (figure 5.2; Ehlers, 1990a). Grains can also show a preferred direction with the long axis parallel to the flow direction, the so called 'fabric'. This makes them very useful for reconstructing old flow patterns of the ice. Besides clast fabrics also shear folds and lamination can occur that indicates the flow direction of the ice. This is called glaciotectonic deformation. The till is commonly overconsolidated due to the weight of the ice cap. Usually, no sedimentary structures are present, however faulting and shearing structures usually occur (Boulton, 1975). Morphologically, a lodgement till surface will generally be characterized by flutes and drumlins (Boulton, 2006).
When a stagnant glacier melts, the debris of the glacier is deposited as melt-out till (Boulton, 1970).

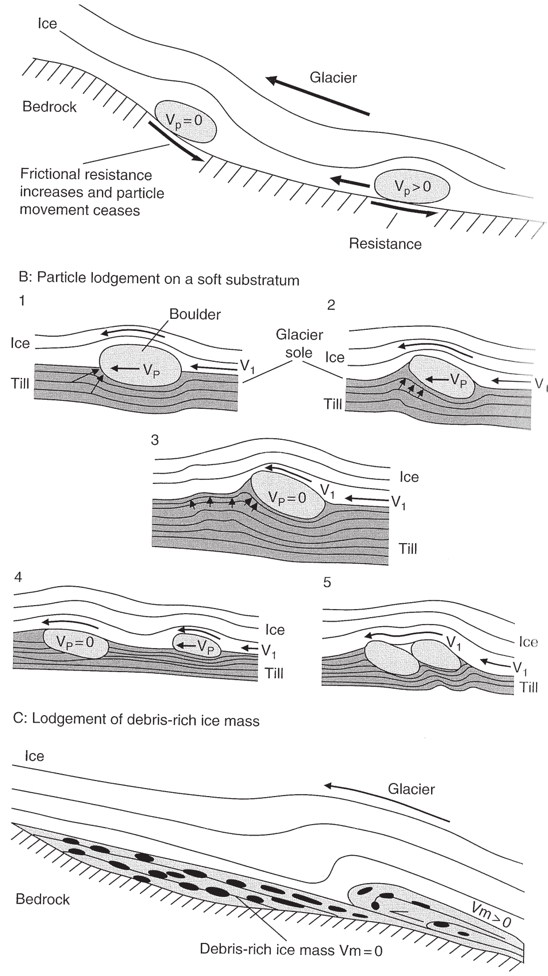


Figure 5.2
Particle lodgement below a glacier: a) on a ridged substratum; b) on a soft substratum; the pebbles or boulders are imbricated in a way they offer the least resistance to the ice flow. Other pebbles may jam against the first and a boulder pavement or concentration can occur. c) lodgement of a debris rich ice mass. From Bennett & Glasser (2009); after Boulton (1982).

Melt-out till can also be formed when the glacier is still slowly moving (Benn & Evans, 1998). It can form on top of the glacier (as supraglacial melt-out till) or underneath the glacier (subglacial melt-out till). Like lodgement tills, subglacial melt-out tills show rounded spherical clasts, although less pronounced. Usually, the original fabric is still dominantly present although it can be disturbed during the melt-out process depending on the conditions of water flow. This may cause some sediment sorting.
Deformation till may form under a glacier due to the shearing of the glacier on its bed. When the original, mostly soft, sediments under the glacier are heavily disturbed it is referred to as a deformation till (Ehlers, 1996). The style of glaciotectonic deformation can range from minor faulting to tectonic lamination and a homogenized diamicton (figure 5.3; Hart & Boulton, 1991). Because, they are not directly deposited by the ice Stephan & Ehlers (1983) call them 'subglacially disturbed sediments'. Like lodgement tills, deformation tills commonly occur in drumlins and flutes (Boulton, 2006).

Supraglacial till
Supraglacial till comprises the formerly distinguished supraglacial melt-out tills (figure 5.4) and flow tills. In supraglacial melt-out tills very angular and non-spherical clasts are present and the grain size

distribution is commonly coarse and unimodal, the fabric is usually not related to the ice flow.

Because the ice does not melt at the same rate everywhere an irregular topography is formed, causing a lot of slumping. This finally generates a irregular topography when all the ice has melted, the result may be a hummocky moraines (Boulton, 2006).

Melt-out till deposits of continental glaciers are usually not very thick because their thickness is limited by the mass of debris is the total column of ice (Boulton, 2006). In the case of large ice sheets, most material is transported at the base of the glacier (Bennett & Glasser, 2009). These two facts will cause the average thickness of melt-out tills after deglaciation in the research area to be relatively low. Besides, its preservation potential is limited as it is positioned on top of the sequence and it is not over-consolidated. This causes the till to be sensitive for reworking by new glacial advances or mass movements, like solifluction (Ehlers, 1996).

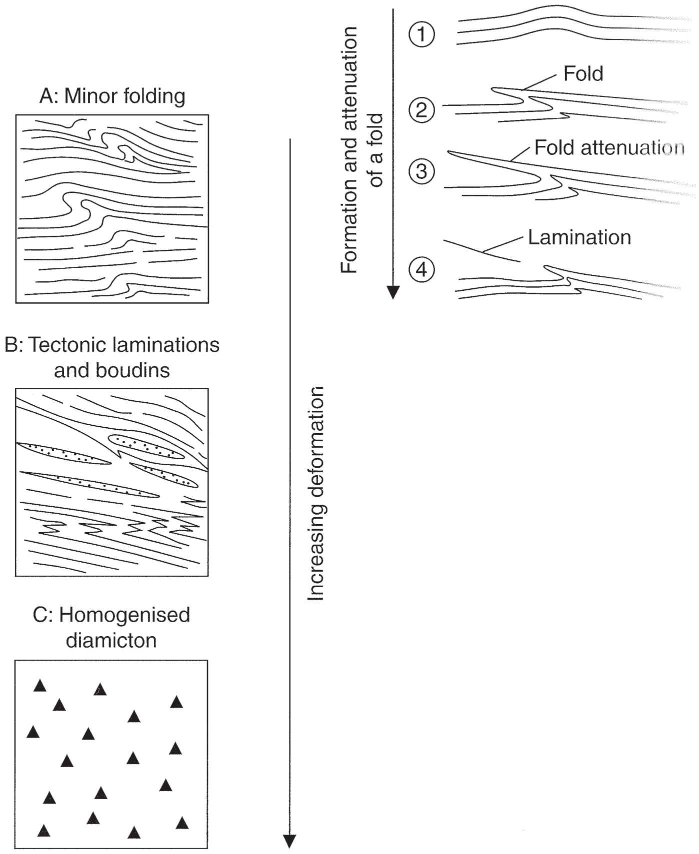


Figure 5.3
Different levels of subglacial deformation and the formation or tectonic lamination from folds. (from: Bennett & Glasser, 2009; after Hart & Boulton, 1991).

Flow till is a secondary supraglacial till (Ehlers, 1996), that has been redeposited by a solifluction-like process. The debris on top of the glacier can become saturated with water and a debris flow can occur. The debris characteristics are very similar to supraglacial melt-out till. Except for the fact that flow tills can be sorted by the flow and that sandy an silty layers can occur. On the bottom of the flow an erosional surface may be found, folds may also be present. (Bennett & Glasser, 2009)

Other tills

Subaquatic tills are deposited under water, e.g. in ice dammed lakes. In these tills dropstones and vertically oriented sand grains can be found (Carr, 2001). The presence of marine or lacustrine shells can also indicate subaquaous till. Moreover a subaquaous till is usually not marked by an unconformity at the bottom, unlike below terrestrial tills (Bennett & Glasser, 2009).

5.1.3. Till stratigraphy and the use for reconstructions

An ideal till sequence of one single glaciation event would contain a lodgement till at the base, with probably a deformation till underneath it. On top of this a small amount of subglacial and supraglacial melt-out till is present, which may be covered by flow tills. The presence of meltwater deposits between two tills can indicate two times the advance of an ice front and melting and retreat between these phases. However, when the upper till seems to be a flow till, the sequence represents one single glaciation event (Boulton, 1977). Another sequence may be lodgement till on subaquous till indicating overriding of a proglacial lake. In practice, these tills can be very hard to distinguish and hiatuses may occur due to glacial erosion.

In the past, till layers with different properties were marked as being from another glaciation phase (eg. Zandstra, 1987; 1993; chapter 6.1 & 7.2). The main assumption behind this is that every ice stream carried a distinct erratic and gravel assemblage from a different source area. Because many other factors contribute to the till composition (chapter 4.5), the view that tills of one single glaciation must be homogeneous is nowadays abandoned. Till composition is only considered to be a reflection of the source area. Despite this, the stratigraphy of tills can serve as a useful tool in the reconstruction of former ice streams. The distribution of very distinct tills may still serve as a proof for different ice streams from different source areas.

Tills are very useful for reconstructing ice streams, especially when till stratigraphy is combined with fabric analyses and the orientation of shear structures. Valuable information can then be obtained indicating the source of an ice flow and the regional ice flow direction.

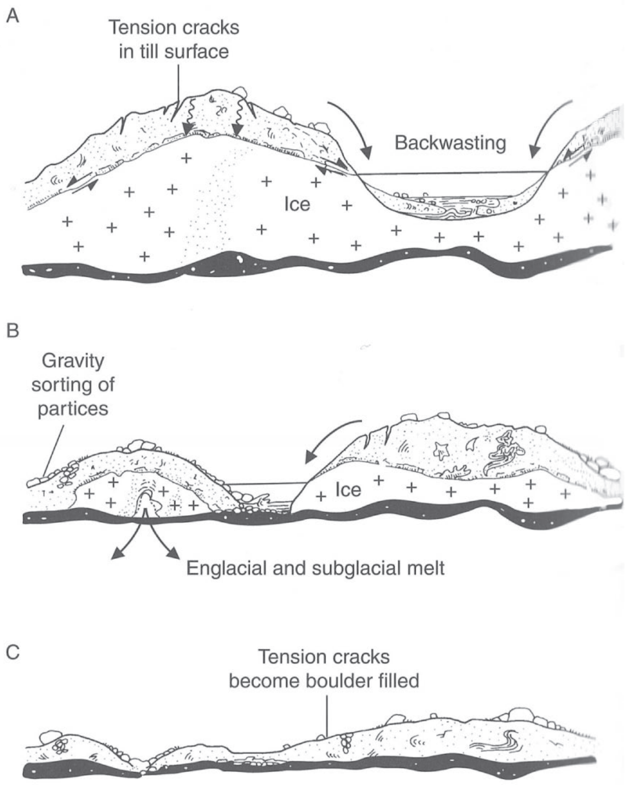


Figure 5.4
The formation of a meltout till. A lot of different processes are involved in reworking of the sediment (from: Bennett & Glasser, 2009)

5.2 Streamlined features and glacial lineations

5.2.1. Drumlins and rogens

Drumlins are smooth, oval-shaped or elliptical subglacial landforms which consist mainly of till. Its long axis is orientated parallel to the ice flow (Clarck et al., 2009), the steepest end indicates the direction the ice flow came from. The length/width ratio is often varies between 1.7 and 4.1 (Clarck et al., 2009). They tend to occur in groups of drumlins fields. Several hypotheses exist about the origin of drumlins (Boulton, 1987; Ehlers, 1996). The first hypothesis assumes that subglacial meltwater plays an important role. Basal ice flow eroded the glacier to form cavities. In these cavities glaciofluvial deposits were deposited. The second hypothesis assumes a direct glaciogenic origin either by erosion or by deposition formed by a spiral-shaped sediment transport of mainly lodgement till. Drumlins tend to occur relatively close to the ice margin.

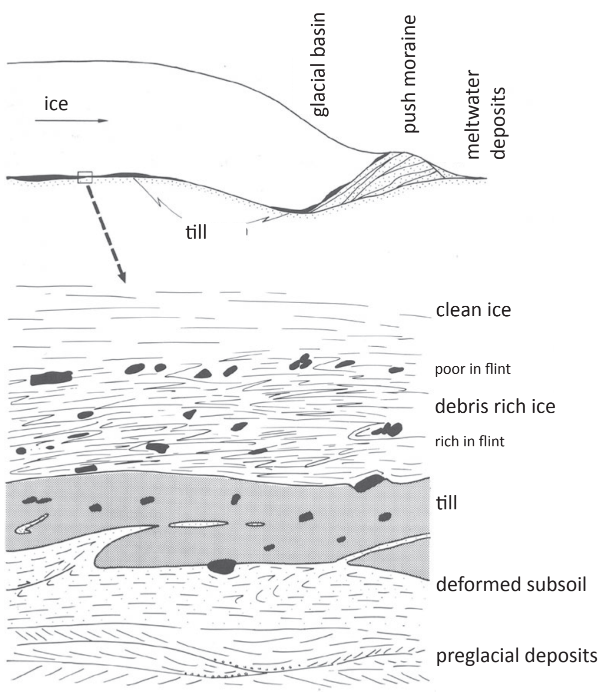


Figure 5.5
The sequence of till composition in the Netherlands. The debris poor in flint is deposited on top of the flint rich debris, because the source area of the latter is further away. Vertical scale: ~ 10m. After, Rappol, 1991

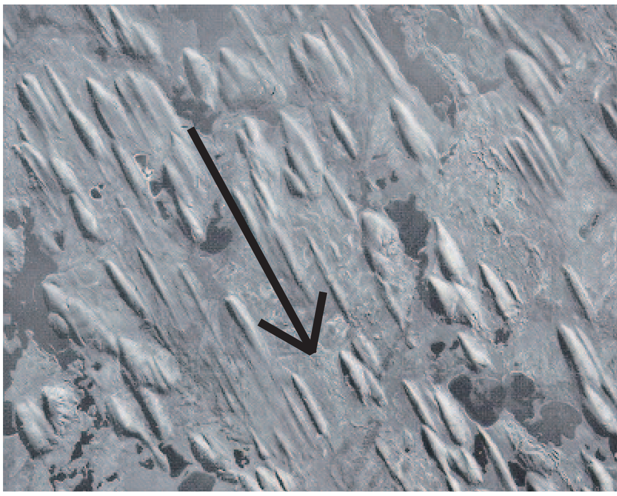



Figure 5.6 (left) Hillshade image of a drumlin field in Canada, the arrow indicates the palaeo-flow direction.
Figure 5.7 (right) Small scale flutes on Bruarjokull, Iceland.

Rogens are drumlinised ridges that are transverse in the direction of the flow which tend to occur close to ice sheets centres (Boulton, 1987). They often have a lunate shape in plan view with the concave shape pointing in a down-ice direction. They have not been found in the research area.

5.2.2. Flutes and megaflutes

Flutes are long straight glacial lineations that form parallel to the last ice flow direction (Boulton, 1976). Usually, they are not higher than 2 meters, but they can be 50 m wide and several hundreds of meters long. They are typically composed of tills and formed behind an obstacle. The pressure on the sediment on either side of the obstacle will be higher, causing a pressure gradient that favours the formation of a small straight ridge. This relief may be accentuated by glaciofluvial activity. Although this theory is widely accepted it is still not certain if this is completely correct. Megaflutes are higher (>5m) than flutes and have a length/width ratio of more than 50 and are therefore more commonly preserved in old glaciated areas, in the research area they are common on the till plateaus.

5.2.3. The use for reconstructions

All these features directly indicate the paleo-ice flow direction of the ice at the time of formation. Because the ice flow may change within one glaciation on a certain locality, older drumlins or megaflutes

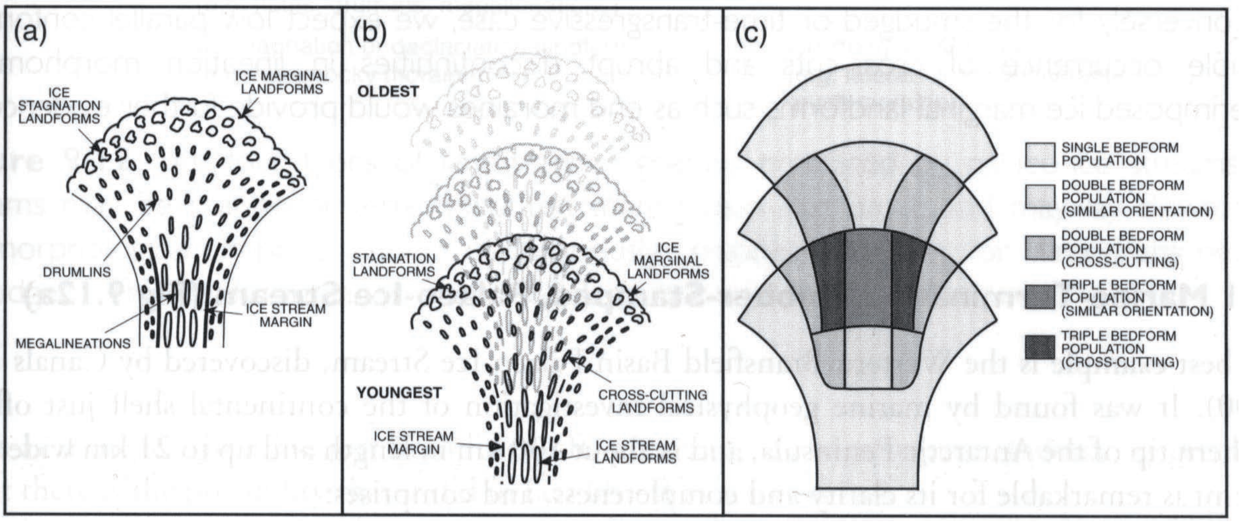


Figure 5.8
Landforms and their cross-cutting relations that can be formed when an ice stream readvances during retreat. From: Clarck & Stokes (2003).

may be partially of totally reworked. Therefore these features primarily indicate the youngest flow direction on a certain locality. However, earlier formed features are commonly partly preserved yielding older and younger features to be superimposed on each other (figure 5.8). The superposition of features, cross-cutting relations and imprinting relations of the lineations gives relative ages of ice flows and their paleo-flow direction. ice-pushed ridges may also be affected by a younger ice flow that causes its surface to be drumlinized or fluted, this happened in the Northern Netherlands (chapter 6.2). The patterns of ice flow chronology can be obtained from satellite images and elevation maps and subsequently used for paleo-reconstruction as described in chapter 2.2.

5.3 Ice-pushed ridges

Ice-pushed ridges (Dutch: stuwwallen, German: Stauchrücken or Stauchwälle) form at the margins of the ice sheet by the pushing up of the deformed subsoil by the ice. Other terms that are frequently used are glaciotectonic ice- or sub-marginal moraines, thrust moraines and push moraines (Bennett, 2001). Here, the term ice-pushed ridge is used, as the ice-pushed ridges in the study area are mainly composed of pushed up preglacial substrate instead of tills. Large composite ice-pushed ridges, which are treated here, form as a result of a positive mass balance or an ice stream. Usually, ice-pushed ridges consist of the substrate beneath the glacier, but tills and meltwater deposits are also very common.

5.3.1 Factors involved in the formation of ice-pushed ridges

The occurrence of ice-pushed ridges is spatially restricted; they do not occur along every ice margin. Therefore, they are associated with specific glaciological and specific substrate conditions, which in combination favour deformation of the glacial foreland (Bennett, 2001). The factors involved in the formation of ice-pushed ridges are outlined below, compiled from Van der Wateren (1995), Bennett (2001) and Bakker (2006).

Structure of the pre-glacial substrate and basement

Main features of the preglacial substrate and basement in the study area include: tectonic faults, folds, salt domes and the basin tectonic dip. These substrate characteristics influence the regional hydrology which appears to affect subglacial resistance conditions. Generally, the Tertiary strata are finer grained and less permeable that the overlying strata. The structures cause older strata with specific hydrological properties to be present at higher positions. The updoming of salt domes, for example, causes the deeper Tertiary clay layers to appear at shallower depths. They can subsequently act as a décollement (figure 5.9), for example in the northern part of Lower Saxony (Kuster & Meyer, 1979). In the western and central Netherlands the ice-pushed ridges line up with the Roer Valley Graben shoulder fault system (Van Balen et al., 2005). In adjacent Germany the Rehburg line is positioned at the very margin of the Lower Saxony Basin, just north of outcropping Mesozoic rocks (Van der Wateren, 1995; 2003).

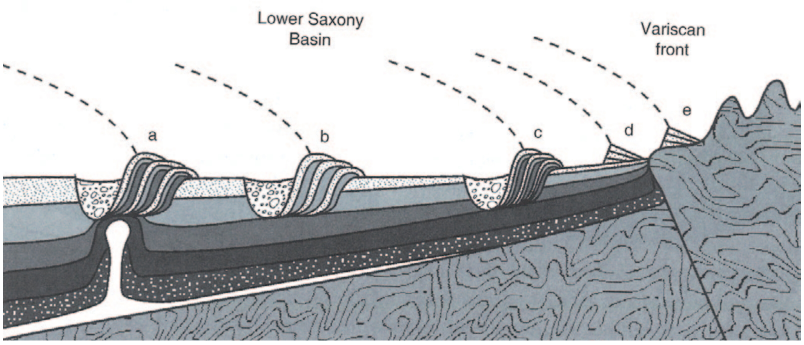


Figure 5.9
Schematic profile of the German Variscan highlands (Weserbergland) showing the association of substrate geological structure and ice-pushed ridges. a) A salt dome brings up clayey layers providing a suitable décollement, b, c) Shallowing of the basin provides a décollement. d,e) Tills and outwash plains develop instead of ice-pushed ridges due to the absence of a suitable décollement. From: Van der Wateren, (2003).

Strength, rheology and hydrology of the foreland

The interaction of glaciers with the foreland of a sedimentary basin involves aspects of subglacial hydrology, favouring the formation of ice-pushed ridges. The foreland deformation can be ductile ,causing folding of strata, or brittle, causing break up into distinct slabs. Whether materials are deformed in a ductile of brittle way depends on the grain size and hydrological properties of the material and on the rate at which the stress is exerted. When stress is applied rapidly, materials tend to deform in a brittle way. The ice-pushed complexes in the central Netherlands show that the coarse grained Middle Pleistocene fluvial deposits have acted as ridged blocks. The explanation is that meltwater was able to drain immediately, causing brittle deformation. A second effect of the coarse material is that it increases the basal friction of the ice sheet (Boulton, 1986; Van den Berg & Beets, 1987). This caused a thickening of ice which overrides coarse sand. Lastly, the basal meltwater production is believed to increase causing more subglacial erosion and piping (Boulton & Hindmarsh, 1987). The latter process may have caused the separation of the ice front in the central parts of the Netherlands into lobes (Van den Berg & Beets, 1987).

Décollement

Clayey and loamy layers have high porewater pressures and can therefore serve as a gliding plane for thrust sheets and nappes (Van der Wateren, 1995). In the eastern part of the Netherlands and in Germany Tertiary marine clays or Pleistocene fluvial clays mainly served as a décollement. The presence of a décollement is directly related to the substrate architecture (figure 5.9), whereas the thickness of the thrusts is determined by the regional contacts between coarser and finer grained strata (Van den Berg & Beets, 1987)

Permafrost

Some controversy exists on the nature and extent of permafrost during the Saalian glaciation in the research area. In the 1960's and 70's several authors stated that permafrost was a prerequisite for the formation of thrust sheets to be moved as rigid masses (e.g. Jelgersma & Breeuwer, 1975). De Jong (1955) directly related the thickness of thrust sheets to the depth of the old permafrost level.

On the other hand, Van der Wateren (1985; 1995) and Van den Berg & Beets, (1987) have shown that the presence of permafrost is not a necessity for glacial thrusting and that ponding by glacial meltwater must have hindered the formation of permafrost. Therefore, Van der Wateren (1985; 1995) states that permafrost rather hinders than favours the formation of thrust sheets.

These results contradict research done by Boulton (1999) in Spitsbergen. He concluded that the permafrost does control the thickness of thrust sheets in modern ice-pushed ridges. Bakker (2006) noted that the composition of the ice-pushed ridges in Spitsbergen differs from the Dutch ridges, the latter contain coarser grained material, yielding different tectonical and hydrological properties. Waller et al. (2009) stated from field studies in the UK, that permafrost at ice margins in mid-latitude areas was much more involved in glaciotectonics than previously thought. From the range of observations available, Bennett (2001) concludes that permafrost is an essential pre-requisite for some types of ice-pushed ridge, but not necessarily for all. He considers the absence of a clear décollement layer to be indicative for permafrost presence during the formation of the ice-pushed ridge.

Preglacial relief

Small differences in pre-glacial relief in combination with high pore water pressures may favour the transfer of stresses to the subsurface and cause pushing of the subsoil (Bakker, 2006). These differences may include outwash fans built up near ice margins (Boulton, 1986) or micro relief in river terraces of the early and Middle Pleistocene rivers (Banham, 1975). As the preglacial reconstruction of the relief in large parts of the research area is difficult, Van der Wateren (1985) concluded that the pre-glacial relief was not important relative to the ice thickness.

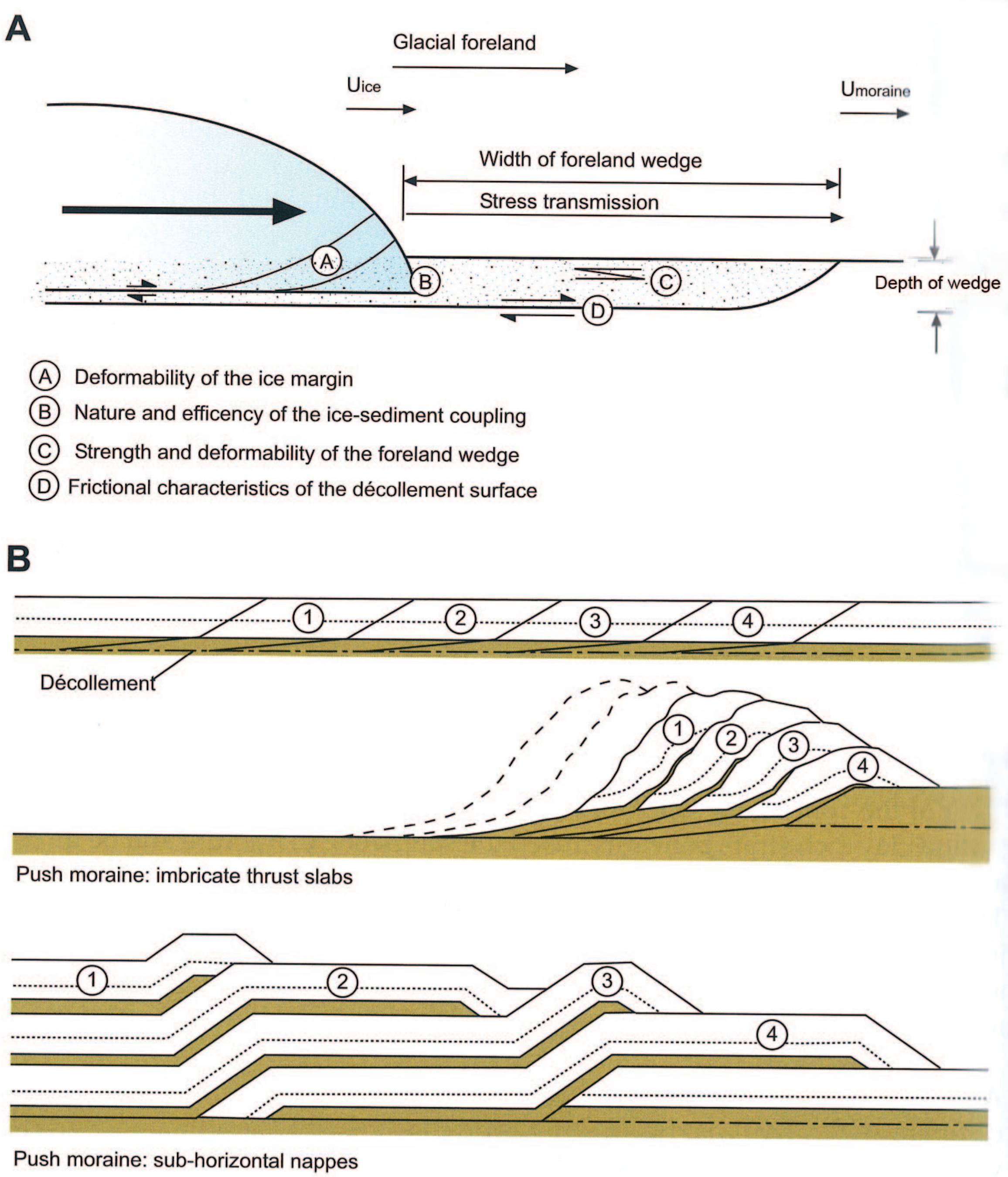


Figure 5.10
The anatomy of a ice-pushed ridge. a) Some basic definitions and key variables. b) Geometry of an imbricate ice-pushed ridge compared to one built of nappes. From: Bennett & Glasser (2009); after: Van der Wateren (1995).

Glacier dynamics
An advance or re-advance of an ice sheet is a prerequisite for the formation of ice-pushed ridges. It occurs due to a positive mass balance in the ice sheet, or because of a local surge or ice stream. The glacier dynamics are closely linked to the resistance of the substratum which allows slow or fast ice flow.
The classical idea about the formation of ice-pushed ridges is frontal pushing of strata by the force generated by the ice movement itself, i.e. the 'bulldozer effect'. Van der Wateren (1985; 1995) demonstrated that this model is not correct because the resistance of a ridge is too high compared to the basal shear stress exerted by the glacier to allow the formation of ice-pushed ridges. Instead the lateral pressure generated by the glacier is more important (Aber et al., 1989). This gradient is generated

by the steep ice sheet profile at the margin that creates a differential gravitational loading. When this ice sheet is actively growing at the margins thrusting occurs, unconsolidated weak sediments are most susceptible to this mechanism (Van der Wateren, 1985; 1995). In reality, both these precesses seem to play a role in the formation of ice-pushed ridges (Ehlers, 1996; 2005).

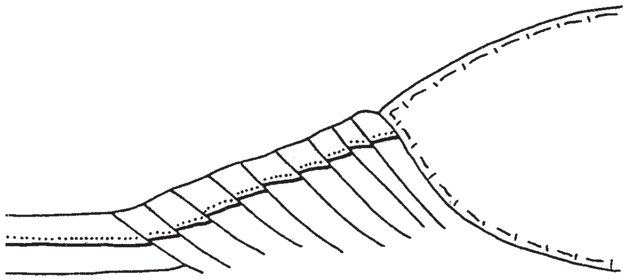


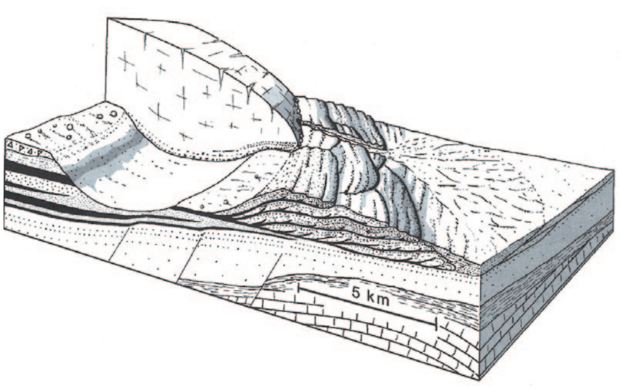
Figure 5.11
Schematic drawing of 'blocktectonics' in a ice-pushed ridge, from: Rappol (1993a).

5.3.2. Glaciotectonic styles in ice-pushed ridges

In ice-pushed ridges sediments can be deformed in four different ways.

1. The first style is the formation of glaciotectonical nappes (Dutch: dekbladen; German: ???; Style 'D' cf. Van der Wateren, 1995; figure 14). Nappes are large packages of sediment that are pushed sub-horizontally on top of each other (figure 10 stuwwal_ontstaan2). Above and below the nappes shear zones are present. Successively younger nappes appear to have formed progressively beneath older ones, carrying them forward into the foreland (Van der Wateren, 1995). Nappes can transported over several several hundreds of meters or kilometers. Nappes are found in several ice-pushed ridges in the research area, and for example also in Denmark (Klint & Pedersen, 1995).
2. The second style is the formation of thrust sheets (Dutch: schubben, German: Schuppen, Style 'C' cf. Van der Wateren, 1995; figure 14. Thrust sheets are thin packages of sediment that are, such as nappes, pushed on top of each other figure 10 Thrusts differ from nappes because both the thrust sheets and the sediments within them dip several tens of degrees towards the pushing direction of the ice.
3. Sediments in a ice-pushed ridge can also be folded (style 'B' cf. Van der Wateren, 1995), yielding a sequence of anticlines and synclines.
4. A fourth style is the occurrence of thrusts displaced over a very small distance, called 'blocktectonics'. The small thrusts (figure 11 Stuwwal_typen, C) did not move over a very large distance, but large abundance of the thrusts yields a large cumulative effect (Rappol, 1993a;b). This style of deformation is very rare and was only found at one locality in the research area (Sibculo-Kloosterhaar ice-pushed ridge).

These styles can occur within one single ice-pushed ridge but also within a nappe, folding and thrust



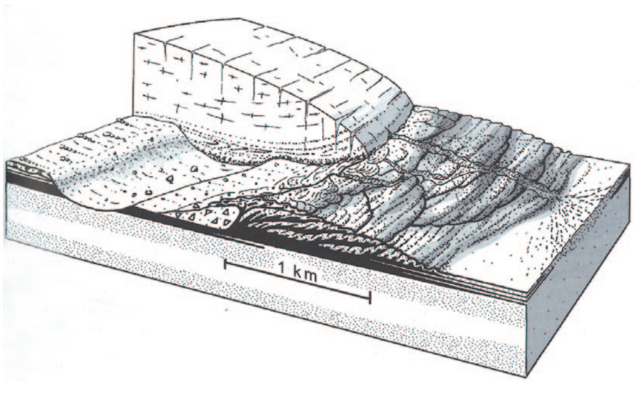


Figure 5.12 (left) Model of a ice-pushed ridge composed of coarse grained brittle deformed sediments. This model is based upon the Dammer Berge in Germany (chaper 6.2), from Van der Wateren (1995; 2003).

Figure 5.13 (right) Model of a ice-pushed ridge composed of fine grained ductile deformed sediments. This model is based upon the Holmströmbreen glacier in Spitsbergen, from Van der Wateren (1995; 2003).

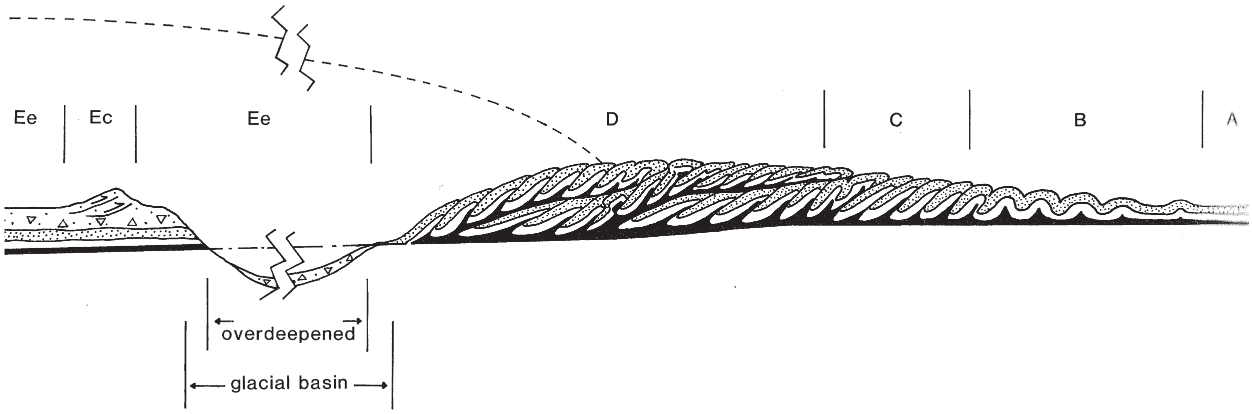


Figure 5.14
Glacialtectonic styles on the ice margin. A undeformed foreland, B steeply inclined structures, C. thrusts and folds, D nappes, E. extensional features, subglacial tills. Ec, compression style, Ee extension style. From: Van der Wateren (1995; 2003)

imbrication may occur, particularly towards the distal extremity (Bennett, 2001). Folding structures within nappes are associated with a ductile deformation style of fine-grained sediments (figure 5.13 Thrust moraine ductile), e.g. clay and rythmic lithostratigraphies such as in lacustrine or marine environments, Tertiary lignites and proglacial deltas (Van der Wateren, 1995). Folds and thrusts (style 'B' and 'C') within relatively thin nappes, are formed in relatively stiff and brittle coarse-grained sediments (figure 5.12 Thrust moraine brittle). This model represents ice-pushed ridges that are formed in thick sequences of coarse-grained fluvial and glaciofluvial deposits, sediments that favour the transmission of stress over great distances. This type of ice-pushed ridge is found in the central Netherlands and the Rehburg line in Germany (chapter 6.2; Van der Wateren, 1995; 2003).

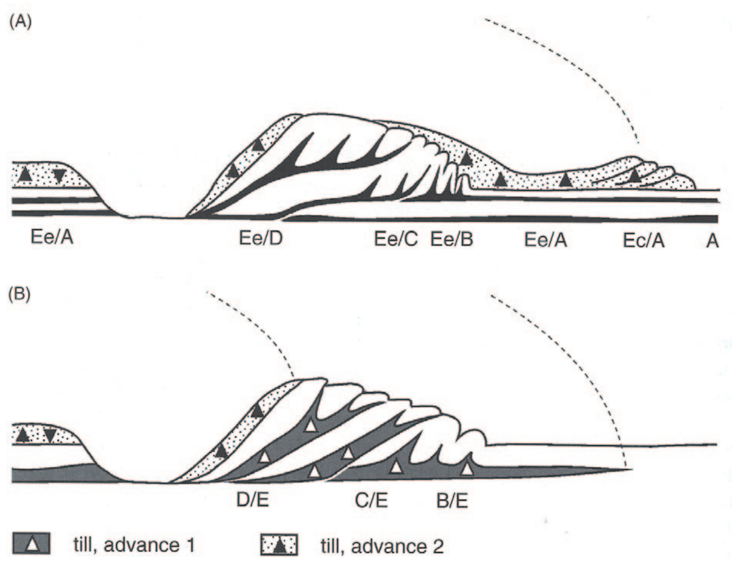


Figure 15 Overprinting of glaciotectonic styles. A) advance sequence, B) readvance sequence in which the older till from advance 1 is incorporated. From: Van der Wateren (2003).

Dump moraines

Another type of (push) end moraine is a dump moraine or ablation moraine. This type mainly consists of tills instead of preglacial pushed materials. They can be formed when the ice front stagnates and pushing can be involved (Ehlers, 1996). In fact, this type of moraine (when well expressed and once containing considerable volume) is indicative for a relatively long stagnation of the ice front (Bennett & Glasser, 2009). Large dump moraines will be formed when much material can be delivered, i.e. with a reasonably fast ice flow while the ice front stagnates, or in places where it contained a relative large amount of debris. Dump moraines are more common in valley glaciers than in continental ice sheets, because the latter contain less englacial debris.

5.3.3. The use for ice sheet reconstructions

Ice-pushed ridges can be very useful in reconstructing Pleistocene glacier landscapes, as their occurrence may provide data on glaciodynamics, ice sheet extension and the palaeo-environment of the glacial foreland (Bennett, 2001). Evidence of overriding or occurrence of 'older' tills and meltwater deposits within a ice-pushed ridge can give implications for the chronology of till/meltwater deposition and pushing (Van der Wateren, 2003) (figure 15 Overprinting of gltecstyles).

A major drawback is the fact that some ice-pushed ridges are controlled more by substratum than by glaciological conditions, which means that alignments of ice-pushed ridges do not necessarily form synchronously (Van der Wateren, 1995). Neither do they represent a stable ice front that is in equilibrium with climate (Evans & Rea, 2003).

5.4 Glaciofluvial landforms and deposits

5.4.1. Tunnel valleys

Underneath an ice cap meltwater can be concentrated in channels. When the flow conditions are energetic and the substrate rather soft, these flows are able to erode the subsurface. Tunnel valleys are most likely formed by subglacial channels with meltwater under a large hydrostatic pressure. Woldstedt (1913) suggests that pressure directly exerted by the ice played a major role in forming the valleys because some of them are u-shaped. Which hypothesis is true is hard to say because insufficient data is available on the morphology and distribution of tunnel valleys (Bennett & Glasser, 2009). Tunnel valleys can be infilled with glaciofluvial sediments, so called N eskers. Glaciolacustrine deposits are also commonly formed during the deglaciation. The most striking tunnel valleys were formed in the Elsterian (chapter 3.5.2). During the Saalian their formation was rather restricted (Ehlers, 1996).

5.4.2. Eskers

Instead of severely eroding the subsoil, meltwater streams underneath the ice can also melt the glacier itself form R-channels. Then an elongated ridge (figure 5.16) up to tens of kilometers long, composed of glaciofluvial sediments is formed, called eskers. They may form when the subsoil is unable to drain all the meltwater (Boulton, 2006; Boulton et al., 2009).

Eskers usually form underneath the glacier surface, but englacial and supraglacial eskers also exist. The glaciofluvial sediment often consists of coarse grained sand and gravel, boulders can also be found. The sediment is often well-rounded (Ehlers, 1996). The subglacial drainage is influenced by the inclination of the ice surface, rather then the bed topography (Flint, 1971). Eskers can only be preserved when the ice has ceased to move (i.e. 'dead ice' conditions), otherwise they would be mainly reworked. Therefore the course of an esker usually reflects the last flow direction of the ice, which makes them useful for paleo-reconstructions (Ehlers, 1996).

Figure 5.16
Esker preserved as an elongated ridge in Canada, source: Geological Survey Canada

5.4.3. Sandurs

Sandurs are outwash plains that are formed at the front of a stationary ice margin (figure 5.17). The apex is located at the point where meltwater leaves the glacier, this can be the end of a subglacial channel. The channels of the sandur are

Figure 5.17
Sandur in Alaska, source: USGS

braided and the sandur has a hyperbolic form in the length profile. Sandur deposits are glaciofluvial deposits and contain both fine sand and coarse gravel. The coarsest material occurs close to the apex. Coarse material can also be found further away form the apex as a consequence of the transport and melting of large ice blocks containing coarse sediments. Besides, a seasonal peaks of meltwater input yields deposition of coarser material. Coarse material will also be deposited when a catastrophic event, a so called jökulhaup, takes place. Jökulhaups can take place when a volcanic eruption occurs or when an ice dammed lake suddenly drains (Bennett & Glasser, 2009).
Large sandur complexes develop when an ice front is stationary. These complexes have interlobating and coalescing sandurs along the ice front. When the ice front retreats, large and continuous sandur complexes may serve as a barrier for the melt water deposits and a lake can be formed. However, ice-pushed ridges are more effective as a water barrier, because of their low permeability. The volume of sandur deposits may give an indication for the time of stagnation of the ice front. OSL-dates from sandurs can provide reliable information about when the ice front stagnated.

5.4.4. Ice-marginal rivers and intramarginal rivers

Ice-marginal rivers (Polish: pradolina; German: Urstromtal, Dutch: Oerstroomdal) transported large amounts of meltwater in front of the ice front, along the ice margin forming broad river beds. Ice marginal rivers are quite common in northern Europe compared to America and the Alpine region (Ehlers, 1996). During the deglaciation meltwater rivers were routed to positions behind the former ice front, so called intramarginal melt water rivers. Some of these valleys were able to deeply incise in the subsurface due to the low sea level. Others terminate in temporary lakes and deposited at relatively high elevation.
Ice-marginal rivers and intramarginal rivers usually are braided river systems, therefore they can form wide terraces and their deposits often contain coarse sand and gravel. Most Saalian ice-marginal rivers that formed during the deglaciation have been reoccupied in the Weichselian and can still be traced in the landscape. During the onset of the Saalian glaciation ice marginal rivers must also have existed, as the north flowing rivers were deflected towards the west and all the meltwater had to be drained. Probably, these rivers influenced the ice margins with their latent heat causing permafrost to be severely reduced in the proglacial situation. Another consequence of the presence of these rivers was the flattening of the proglacial relief. Both the absence of permafrost and the state of the proglacial relief had important implications for the formation of ice-pushed ridges (chapter 5.3.1) Most of these deposits have been severely eroded by the ice or deglaciation rivers.

5.5 Glaciolacustrine landforms and deposits

5.5.1 Extra-marginal (proglacial) lakes

Ice margins can block rivers and cause meltwater to accumulate, forming proglacial lakes. The term proglacial lake can be used for lakes that are formed in front of an ice margin or lakes that are strongly influenced by glacial meltwater, not necessary in contact with the ice (Teller, 2003). Extra-marginal lake refers to a proglacial lake formed in front of the maximum extension of the ice, in contrast to proglacial lakes that exist during deglaciation in former ice-excataved areas (chapter 5.5.2). They can be formed when an advancing ice sheet, a ice-pushed ridge or another obstacle blocks a meltwater stream or former river course. The extent and depth of these lakes is controlled by the location of the ice margin, topography of the landscape, differential isostatic rebound and nature of the sediment supply (Teller, 1987). Sometimes these lakes are interconneced by channels or direct overspill. A series of lakes of which a lake drains into the next one, which subsequentely drains into the next is called a cascade. Large proglacial lakes are known from the last glaciation in the Baltic region (Zelčs & Markots, 2004), in Russia (Mangerud et al., 2004) and also in America (Teller, 2003). The presence of proglacial lakes implies locations for the ice lobes that form obstacles behind which the meltwater can accumulate. For example, the presence of the Cleaver Lake indicates the coalescence of the British and the Fennoscandian Ice Sheet. Proglacial lakes can prevent permafrost to establish over large distances and thus they can control glacial advances (Van der Wateren, 2003). They can also trigger ice streams (chapter 4.6).

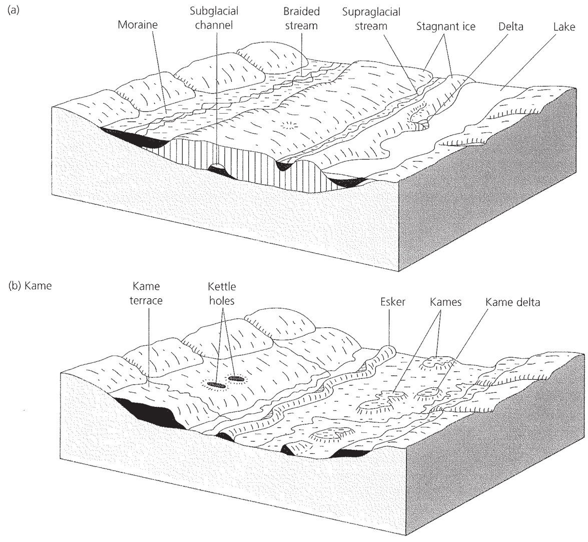


Figure 5.18
Fluvioglacial landformes that form during a) the glaciation and b) the deglaciation. After: Flint, 1971

The morphological name for lake deposits that formed in direct contact with the ice front (extramarginal) is kame terrace. When the ice melts, part of the marginal lake deposits can stay intact, the ice-contacted edges can be subject to a lot of folding and faulting because of the melting of the ice. Other characteristic features of kames and kame terraces are kettle holes, depressions that form as a result of the melting of a large ice block (figure 5.18). After melting of the ice, they occur as irregular hills and ridges in the landscape.

Also, glacio-deltaic sediments can be deposited in subaquous fans by meltwater tributaries. Near the apex sands and gravel were deposited in the foresets. Further away in the almost horizontal bottom sets fine sand, silt and clay could be deposited. In the fine sediments varves are common. This annual layering is caused by more input of sand in the summer and more input of clay and organic material in winter. Dropstones are very common. The sequence of these sediments can give an indication for the proximity of the ice front. A delta from glacial meltwater steams can form under both water level rising and water level falling conditions. Also, subaqueous fans can occur where meltwater streams from the ice sheet entered. Sedimentological analysis of these deltas can reveal a lot about lake level chronology. This information can then be coupled to glacier dynamics or melting (chapter 6.6.1 - Winsemann et al., 2009).

5.5.2 Intra-marginal (deglaciation) lakes

Intra-marginal lakes form between the former maximal extension of the ice and the retreated ice front, i.e. during deglaciation. They can form between all kinds of obstacles (ice-pushed ridges, glacial basins, outcropping hardrocks or dead ice masses) or in glacial basins or depressions. During deglaciation large ice-marginal lake systems developed forming complex interconnected systems. Like with extra-marginal lakes cascades can occur. In these lakes lacustrine deposits can accumulate, which may be preserved as kames when the ice retreats. Kames are landforms consisting of meltwater deposits accumulated in the ice or adjacent to an ice margin (Ehlers, 1996). Kames can contain both glaciofluvial and glaciolacustrine deposits and they are usually formed inside the ice sheet (intramarginal).

In the Scandinavian sea both the ice margin and isostasy helped to create a large proglacial lake after the last glaciation, which was eventually replaced by the Baltic sea (Björk, 1995). In the research area these lakes developed mainly in de glacial basins (chapter 6.7), which are filled with varved glaciolacustrine fine grained deposits, which can provide important chronological information.

6. Glacial, proglacial and deglacial sediments and landforms in the research area

6.1 Tills and till plateaus

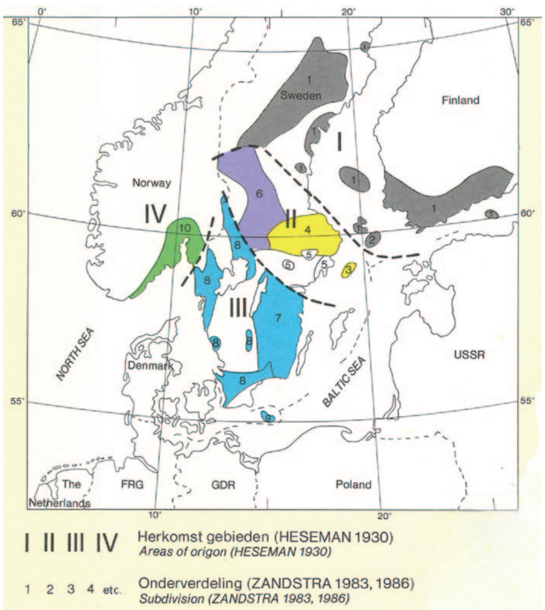
Tills occur very frequently in the research area. Most tills can be found on the till plateaus (German: Geeste), but they can also be traced in glacial basins and be covered by younger deposits. Till plateaus do not only consist of tills, preglacial deposits or proglacial and deglaciation meltwater deposits can also be abundant (figure 6.6 & 6.10).

6.1.1 Erratics and till types in the research area

In Germany tills are referred to as Grundmoraine, Geschiebelehm and in the Netherlands as 'grondmorene' and 'keileem'. In the Dutch lithostratigraphy, tills in are referred to as the Gieten Member of the Drente Formation (Westerhoff et al., 2003; Bakker et al., 2003). A sandy till facies is known positioned on top of the 'common' tills. (Dutch: keizand; German: Geschiebesand). It is called the Gasselte Bed within the Drente Formation (Bakker et al., 2003) and considered as an erosional remnant of till (Edelman & Maarleveld, 1958; Ter Wee, 1979). In this residual till the coarser fraction remained. This till remnant may also be interpreted as originally sandy (melt-out) that was mainly eroded by polar desert deflation, i.e. it has a different composition than the loamy till below it.
Stephan & Ehlers (1983) suggest that most of the till in Lower Saxony has been deposited as melt-out till. For the Dutch situation, this is contradicted by Van den Berg & Beets (1987), Rappol (1987) and Boulton (2006) who conclude that the tills are mainly lodgement tills. The latter is more logical because the amount of englacial debris is relatively low in a continental glacier compared to valley glaciers (chapter 5.1).

Erratics

In the research area a lot of erratics can be found that mainly originate from Scandinavia. These erratics can be found in the (residual) tills, but they are also present in meltwater deposits. The more exact erratic composition reflects the origin of the ice lobe. Hesemann (1930) distinguished four areas within Scandinavia that contained clearly recognizable bedrock composition reflected in Dutch erratics.
With this concept the erratic composition can be shown with a Hesemann formula. This formula consists of four numbers representing the rounded off percentage of occurrence of erratics from



Heseman (1930)		Zandstra (1983; 1986; 1987; 1993)	
I	Eastern Scandinavia	1	Eastern Fennoscandinavia
		2	Baltic Sea, south of Åland
II	Middle Sweden	3	Baltic Sea near Stockholm
		4	Uppland
		5	Stockholm and adjacent area
		6	Dalarna and adjacent area
III	Southern Sweden	7	Småland and Värmland
		8	Southern Sweden
		9	Bornholm
IV	Southern Norway	10	Southern Norway

Table 6.1

Scandinavian source areas for erratics in the research area.

the four areas. For example, an erratic composition with area I: 62%, area II: 21%, area III: 16% and area IV: 1% gives formula 6220. This concept was extended by Zandstra (1983; 1986; 1987) who distinguished 10 areas. He also demonstrated that the clast associations do not represent different glaciation phases, but were deposited by a sequence of different ice masses within a single glaciation event.

Dutch till classification

Zandstra (1986) classified tills in till groups, mainly based on the origin of the erratics in them. Within these groups till types were distinguished that based on flint content and chalk (Rappol & Kluiving, 1992). (Fig table 1 rappol 1987).

The main till groups after Zandstra (1986; 1993) are:

Voorst group ('Schollenkeileem'; De Waard, 1944), is associated with East Baltic components (Area I), very silty and clayey, very calcareous. The fine grain deposits are derived from fine grained deposits from the Baltic Sea, that were present before the arrival of the ice (Rappol, 1991). Within the Voorst Group the flint poor Voorst Type and an unnamed flint rich till type is distinguished.

Heerenveen group, contains mainly erratics from area III. In the eastern Netherlands the Markelo type is distinguished (Kluiving et al., 1991).

Assen group is a relatively sandy till, also associated with Eastern Baltic erratic components (Area I). Within this group the flint rich Assen type and flint poor Emmen type are distinguished.

Rhenen group. Contains erratics from from area II, eastern part, within this group the flint rich Amersfoort type and flint poor Rhenen type are distinguished.

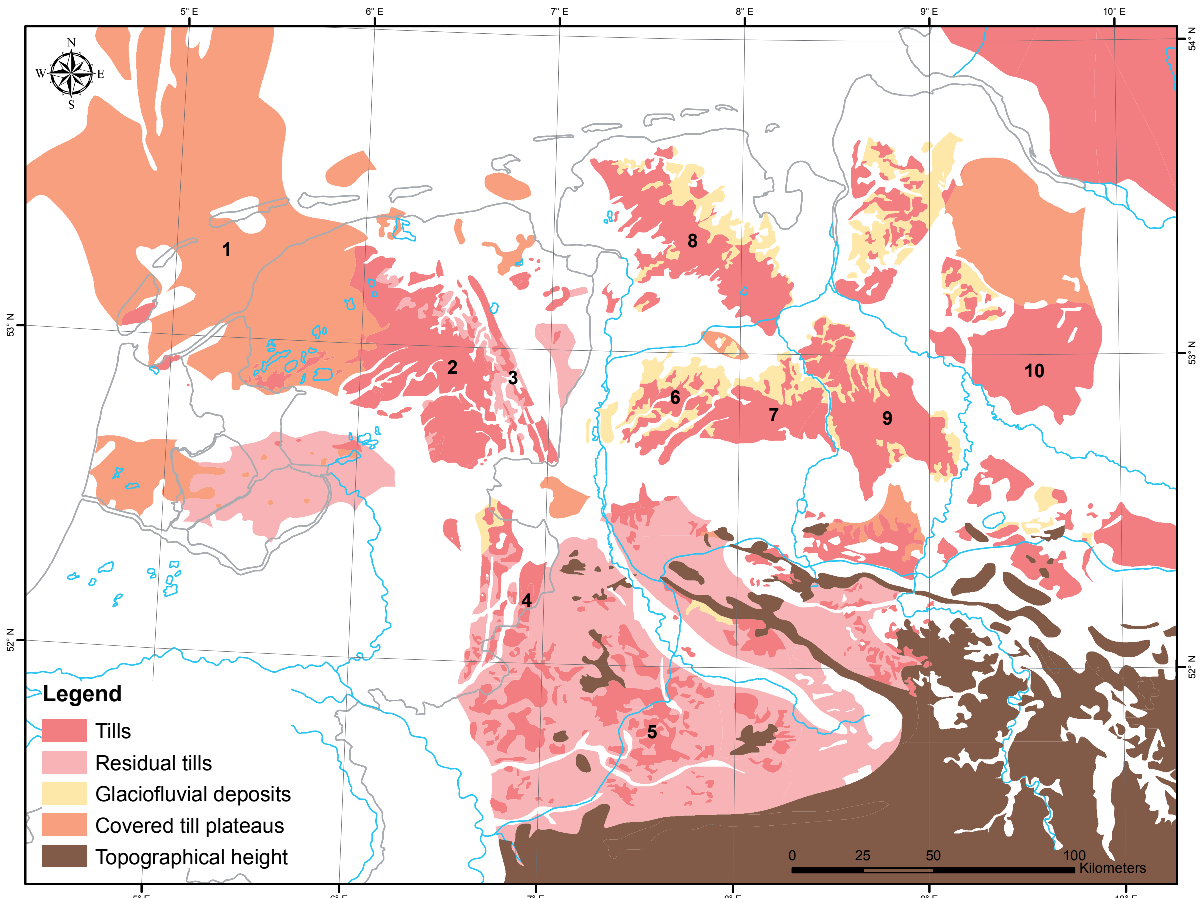


Figure 6.1
Location of the several till plateaus in the research area. 1= North Sea till plateau (Borkumriff Formation), 2= Drenthe till plateau, 3= Hondsrug, 4= Twente-Achterhoek till plateau, 5= Münsterland till plateau, 6= Hümmling, 7= Cloppenburg Geest, 8=Ostfriesland, 9= Syke Geest, 10= Lüneburger Heide.

The flint poor tills are usually found on top the flint rich till, because the flint originated from closer source areas (chapter 4.5). Tills that are rich in chalk are called Geschiebemergel in Germany. Most tills have been decalcified due to chemical weathering. Decalcification is considered to be a secondary feature (Rappol, 1991a), therefore the chalk rich counterparts of the till types (Zandstra, 1990; 1993) are not considered here.
Rappol (1991a) calls these groups the first eastern Baltic till (Voorst group), west Baltic till (Heerenveen group) and the second Baltic till (Assen and Rhenen group) (table 6.1).

Morphology of the till plateaus
Flutings occur very frequently on the till plateaus in the research area. They can still be seen in the landscape today, especially on high resolution digital elevation images. The orientation of the flutings affected the post-glacial river valleys, this was already noted by Edelman & Maarleveld (1958). Ehlers & Stephan (1983) used these forms together with till fabrics to reconstruct ice flows. Today, also high resolution data is accessible to analyse these landforms (chapter 2.2).
During deglaciation large parts of till plateaus were eroded by incising deglaciation rivers. Later on more erosion took place ice marginal rivers in the Weichselian and by local rivers. Despite the substantial erosion some large till plateaus are still present in the research area.

6.1.2 North sea

West and north of the Frisian Islands a calcareous, sandy loam with pebbles was mapped by Joon et al. (1990). They include it in the Borkumriff Formation and interpreted it as the continuation of the gently westward dipping Drenthe Till Plateau (Laban & Van der Meer, 2004). As on the Drenthe Till Plateau, the thickness of the till varies from 1 meter to 10 meter. The tills were found at a depth of 20 meters near the Frisian Islands, and at a depth of 70 m further to the northwest.
Two facies were distinguished (Laban, 1995). The first facies is a sandy till with west Scandinavian gravel. The second facies occurs in the eastern part of the Dutch North Sea and is more gravelly, consisting of greenish-grey, sandy and up to 47% matrix supported Scandinavian gravel. West of Texel some erratics have been found that lie on a bed of aeolian fine sand (Cameron et al., 1986). NW of Texel also NE-SW stretching linear gravel ridges have been found with a similar orientation as the ridges in Drenthe. These tills only stretch out some 50 kilometres into the sea. Further to the west only aeolian and glaciolacustrine sediments have been found. In the central and northern part of the North Sea, Saalian till is also missing. This either indicates that there was no contact between the British and the continental ice sheet (Long et al., 1988) or that the till was eroded after deposition. In the German Bight also till (residual) deposits were found. NE of Helgoland, east of the submerged Elbe ice marginal valley (Figge, 1983).

6.1.3 Drenthe-Friesland till plateau (Northern Netherlands)

In the northern Netherlands the Drenthe-Friesland till sheet is present of 2 to 8 meters thick, to the west the till thickness increases. The southeastern boundary of the till plateau is relatively sharp. The eastern part of the till plateau is defined by a very sharp erosional boundary of the Hunze river valley (Edelman & Maarleveld, 1958). To the west the till plateau continues in the North Sea as the Borkumriff Formation. The southern limit of the till plateau is rimmed by ice-pushed ridges that are mainly composed of tills (chapter 6.2.4) and the Vecht valley (chapter 6.5). The till sheet extents on the western side to Friesland and even the North Sea (figure 6.1).
Essentially one single till sheet covers the Northern Netherlands (Rappol, 1987) with internal differences in composition. The main part of the Northern Netherlands the Heerenveen till group is dominant (Zandstra, 1990; Rappol & Kluiving, 1992), the ice-pushed ridges on the south are characterized by the Voorst till group. On the Hondsrug area the Assen till group is dominant (Zandstra, 1990; Rappol & Kluiving, 1992). The differences in till composition are attributed to different source areas

of adjacent ice lobes, and the composite till sheet is strongly suggested to have formed during one main episode of advance (i.e. no deglaciation between deposition of various layers within the tills) (Meyer, 1987). Also more locally derived sediments can be found in the tills, at the base the tills are more humic because of the presence of reworked middle Pleistocene coastal plain deposits. Several fabric analyses were carried out in this region by Rappol (1984; 1987). They indicate a diverting ice flow, NE-SW in southern Drenthe and NEE-SWW in Friesland and NW Drenthe. This diverting pattern can also be seen very clearly on the AHN (figure 6.2). The ice-pushed ridges near Steenwijk have been fluted in a NE-SW direction. Near Zuidwolde a classical drumlin-like feature can be seen which has been truncated by a NEE-SWW flow on the northern edge. This younger flow occurred between Zuidwolde and the Steenwijk ridges an probably reached the Northern part of the Veluwe where it formed the youngest part of the Veluwe, the Woldberg.

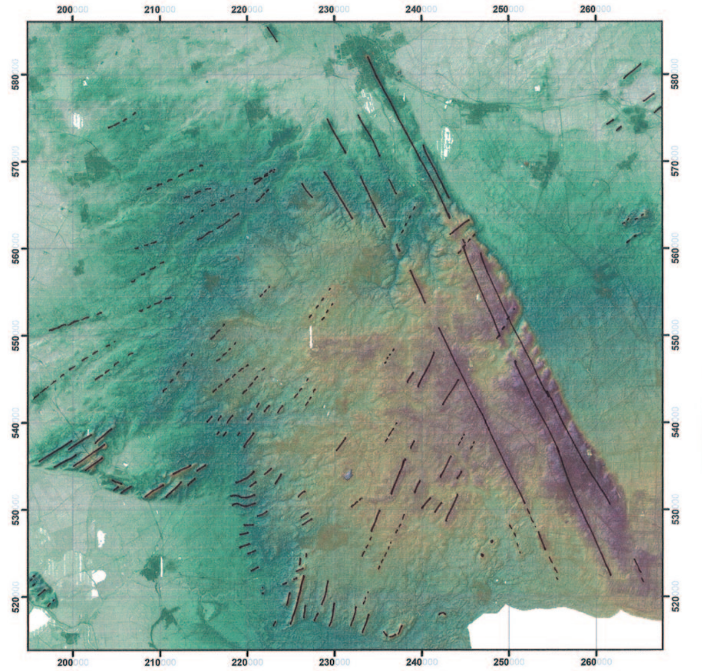


Figure 6.2
AHN image of Drenthe, the glacial ridges are marked by black lines. The Hondsrug lines clearly truncate the earlier NE-Sw oriented ridges.

Hondsrug

The Hondsrug in Drenthe is one of the most prominent geomorphological features in the northern Netherlands. It is an alignment of parallel NNW-SSE running ridges about 50 km long and up to 15 m higher that its surroundings (figure 6.2 Rappol & Kluiving, 1992). It is located at the eastern edge of the Drenthe Till Plateau. Its internal composition consists of tills and preglacial sediments (ref). The erratic assemblage found on the Hondsrug differs from the surrounding areas, Eastern Baltic components are most prominent forming the Assen till group (figure 6.3).

In the past it was suggested that the Hondsrug was formed due to tectonics (e.g. Ter Wee, 1979). However, there are no faults in the Mesozoic and Tertiary subsurface coinciding with the Hondsrug (Van den Berg & Beets, 1987). It was concluded that the Hondsrug was not a neotectonic feature but a megafluted bedform formed by glaciotectonic precesses and marked by glaciofluvial erosion of the Hunze deglaciation river at the eastern edge

Figure 6.3
Detail of the section showing the till of the Assen till group on the Hondsrug. The lower part is the flint rich Assen type and the upper part is the flint poor counterpart: de Emmen type. The photo was taken in the De Boer pit, Emmerschans. From: Ruegg (1983).

(Rappol, 1984; Van den Berg & Beets, 1987). The orientation of these ridges indicate an ice flow that overprints older structures in the Drenthe Till Plateau, some of these older structures can still be seen locally on the figure 6.2. Fabric analysis of the till confirm this flow direction (Rappol, 1984; 1987). Therefore, it was concluded that the Hondsrug represents the youngest NNW-SSE flow direction of the ice (phase 3 in Rappol 1991a). In SE Drenthe NNW-SSE and N-S fabrics and even NNE-SSW directions (figure 6.5).

Figure 6.4
The till section of De Lutte. The three tills can be distinguished based on their color. From: Rappol (1993a)

6.1.4 Twente-Achterhoek-Niederrhein till plateau (Eastern Netherlands)

In the eastern part of Twente and Achterhoek till patches were mapped that probably once formed a more or less continuous till sheet, mainly located on Tertiary deposits.

Near De Lutte, a till section was found that contained three tills. Some of these tills can be found in the whole region. Because this section has been so well described is it outlined here and correlated to the rest of the region. The three tills were found by Kluiving et al. (1991) and Rappol et al. (1991) in the north-south trending Oldenzaal-Enschede ridge (chapter 6.2.2). These tills were distinguished visually based on color and texture (figure 6.4 & 6.5) and correlated to tills in the northern Netherlands (table 6.2). The orientation of the deformation structures were different for the tills indicating different directions of ice flow. The tills contain a lot of local material and probably younger tills contain much reworked material from the older tills (Rappol et al., 1991).

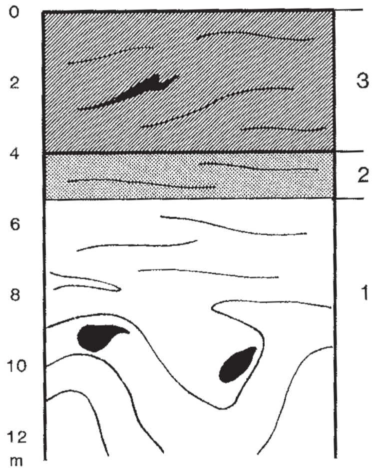


Figure 6.5
Schematic till stratigraphy of the section De Lutte.

Till stratigraphy (Rappol, 1991a)		Till group (Zandstra, 1993)	Till type (Zandstra, 1993)	Northern Netherlands	Eastern Netherlands
3 Second eastern Baltic till		Assen	Emmen, Rhenen*		
			Assen, Amersfoort*		
2 West Baltic till		Heerenveen	Markelo, Heerenveen		
1 First Baltic till		Voorst	Voorst ('schollenkeileem')		
			Unnamed type		

* Occurs in the Gelderse Vallei, central Netherlands.

Table 6.2
Correlation of the tills in the northern and eastern Netherlands. The black tills represent the flint poor till types. After: Rappol et al. (1991).

Till 1: This locally very thick till was interpreted as the flint rich till type of the Voorst till group. Fabric analysis revealed a north-south ice flow. This matches with N-S trending scratches found in a sandstone at Losser by Römer (1978) and a NNW-SSE fabric found near Losser (Rappol et al., 1991). The upper part of till 1 was subsequently deformed by an ice flow to the west (Kluiving et al., 1991). In the Eastern Netherlands, this is probably the only place were this till can be found.

Till 2: On top of till 1, a sandier till was found of 1.5 m thick. This till was heavily reworked and sometimes mixed with till 1. Fabric measurement was very difficult because the till had been deformed too much by the ice flow depositing till 3. Horizontal shear and drag folds however, do indicate a westward ice flow (Kluiving et al., 1991). The same westward flow was found by Rappol (1985) in Markelo. Because of the erratic content and its presumed orientation this till was considered to represent the Markelo till type within the Heerenveen till group (Zandstra, 1986). It was found near Ootmarsum, Oldenzaal, Enschede and further towards the west near Hengelo (Van den Berg & Den Otter, 1993).

Till 3: This 4 m thick till shows a less sandy matrix and contains erratics that predominantly have an East Baltic provenance (Area I). The ice flow direction varies from N-S to NW-SE. This till was correlated to the Assen Till group on the Hondsrug. The till was also found on the Itterbeck-Uelsen ice-pushed ridge (Kluiving et al., 1991; 1994) on the Enschede-Oldenzaal ridge (Van den Berg & Den Otter, 1993) and in the Münsterland Embayment (Zandstra, 1993). The Eastern Baltic erratics are present in this area of Twente, they gradually diminish towards the west (Zandstra, 1987).

6.1.5 Western and central Netherlands

In the western and central part of the Netherlands till plateaus are absent. Till was only found on the bottom of glacial basins at a depth of 30 and 40 meters below present sea level. The till can be as thick as 15 meters. In the Amsterdam glacial basin, tills are found at a depth of 55 m below NAP (De Gans et al., 1987). The erratics in the tills indicate a south Swedish origin (Area III) (Zandstra, 1987; Meyer 1987). In Wieringen, the Voorst till group containing erratics form 'area I' have been found (Rappol, 1991b).

In the Gelderse Valley the sandy Rhenen till group occurs (Zandstra, 1983; Rappol, 1991a). Unlike in other parts of the research area, erratics from the eastern part of area II (area 3 Zandstra) are most dominant (Zandstra, 1987). Till fabric analyses in the Gelderse Valley indicate a NW-SE of NNW-SSE flow (De Waard 1945; Boekschoten & Veenstra, 1967).

6.1.6 Till plateaus in Lower Saxony

In Lower Saxony, several well developed, distinctive till plateaus occur, called Geeste: Lüneburger Heide, Syke, Cloppenburg-Hümmling and Ostfriesland till plateaus (figure 6.1). They are separated by large river valleys. The thickness of the tills varies between 50 m to several tens of meters or even decimeters (Stephan & Ehlers, 1983). During and after deglaciation large rivers developed that eroded large parts of the till plateaus. The fluvial erosion has continued in the Weichselian, when part of the valleys acted as ice-marginal rivers. On the till plateaus local drainage followed the pattern of the glacial lineations.

Till stratigraphy

Till stratigraphy formed the starting point of the phase model developed by Ehlers and coworkers (Ehlers et al., 1984; Ehlers, 1990 - chapter 7.3.1). The till types in Lower Saxony are described here in accordance to their recognized phases.

In North German lithostratigraphy, deposits from the Saalian glaciation are divided into three

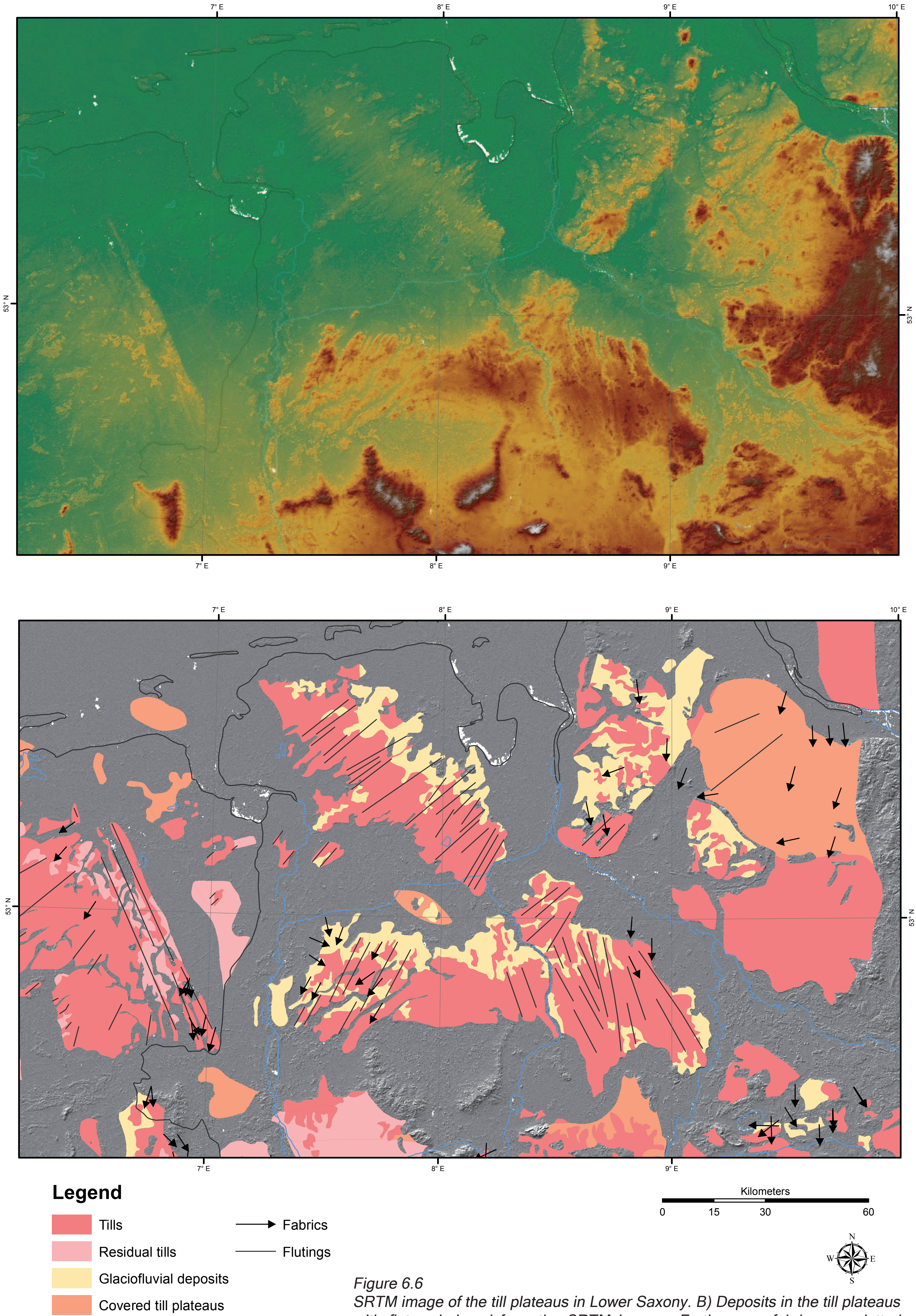


Figure 6.6
SRTM image of the till plateaus in Lower Saxony. B) Deposits in the till plateaus with flutes deduced from the SRTM images. Furthermore fabrics are plotted after Schröder (1978 - Hümmling region); Rappol (1993a - the Netherlands) and Skupin et al. (2003 -Weserbergland) and Ehlers (1990 - northeastern Lower Saxony). For the names of the till plateaus see figure 6.1.

glaciation substages: the Main Drenthian, the Younger Drenthian and the Warthian (Meyer, 1983; 1987); called the older, middle and young Saalian till respectively by Ehlers (1990, 1995, 2005). The main till units are separated by meltwater deposits.

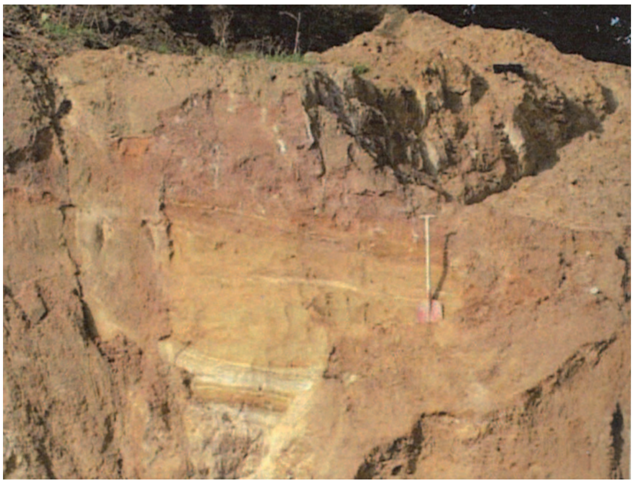


Figure 6.7
Red moraine on the grey brown moraine in an outcrop near Hegel - Cloppenburger Geest (from: Speetzen & Zandstra (2009).

The Main Drenthian advance till: This till was formed during the maximal extension of the ice and covers most of the research area. It is characterized by erratics from south and central Sweden (Meyer, 1983). In Ostfriesland and the Hümmling region erratics from the Oslo region (area 10) are common, which are considered to be originally delivered in the Elsterian (Speetzen & Zandstra, 2009). Locally, this till is overlain by a brown-red clay rich till (figure 6.7) and with an east-Baltic assemblage of erratics (Ehlers et al., 1984; Ehlers, 1990a;b; Caspers et al., 1995; Speetzen & Zandstra, 2009). This occurs especially between the Hunze and Ems rivers, on the eastern side of the Ems and on the western edge of the Hümmling (Speetzen & Zandstra, 2009). They correlate the lower till to the Heerenveen Till group and the red till to the Assen Till group. A similar sequence (figure 6.8) has been found in the Netherlands in the Hondsrug area (Rappol, 1987) and in the Münsterland Embayment (Zandstra, 1993). As in the other till areas, the boundary between these two tills is very sharp and no meltwater of periglacial deposits are present between them. This means that the two tills originate form the same ice sheet and that the area was not ice free between the deposition of the two tills (Meyer, 1983; 1987; Caspers et al., 1995). The stratigraphy and reasoning are similar as adopted in the Netherlands (see chapter 6.1.3 -figure 6.8).

The Younger Drenthe Advance till is a till of 5-10 m thick with more flint and chalk materials (Caspers et al., 1995; Meyer, 2005). Probably, the ice retreat between these phases was very limited (Meyer, 1987). From fabrics it can be concluded that this ice flow came from the NE. This advance reached only the area around Hamburg and Bremen, therefore no counterpart is present in the Netherlands.

The Warthe glaciation till is, like the top part of the main Drenthe till characterized by eastern Baltic tills (Meyer 1987; Caspers et al., 1995; Meyer, 2005). Till fabrics how a main flow from the east and northeast (Stephan & Menke, 1993). – As with the Younger Drenthe Advance there is no Dutch counterpart, since it is retreat phase that occurred around the area of Hamburg.

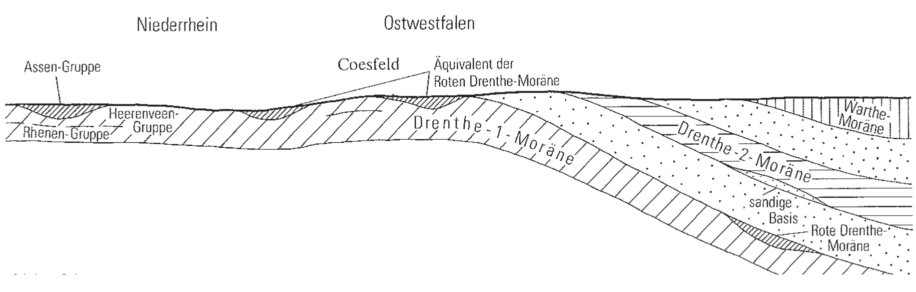


Figure 6.8
Correlation of the tills found in the research area. The Drenthe-1 tills (Heerenveen group) can be found in the whole area, locally overlain by the redbrown Assen till group. In the northeastern part of the research area younger tills occur, separated by meltwater deposits (from: Klostermann, 1992)

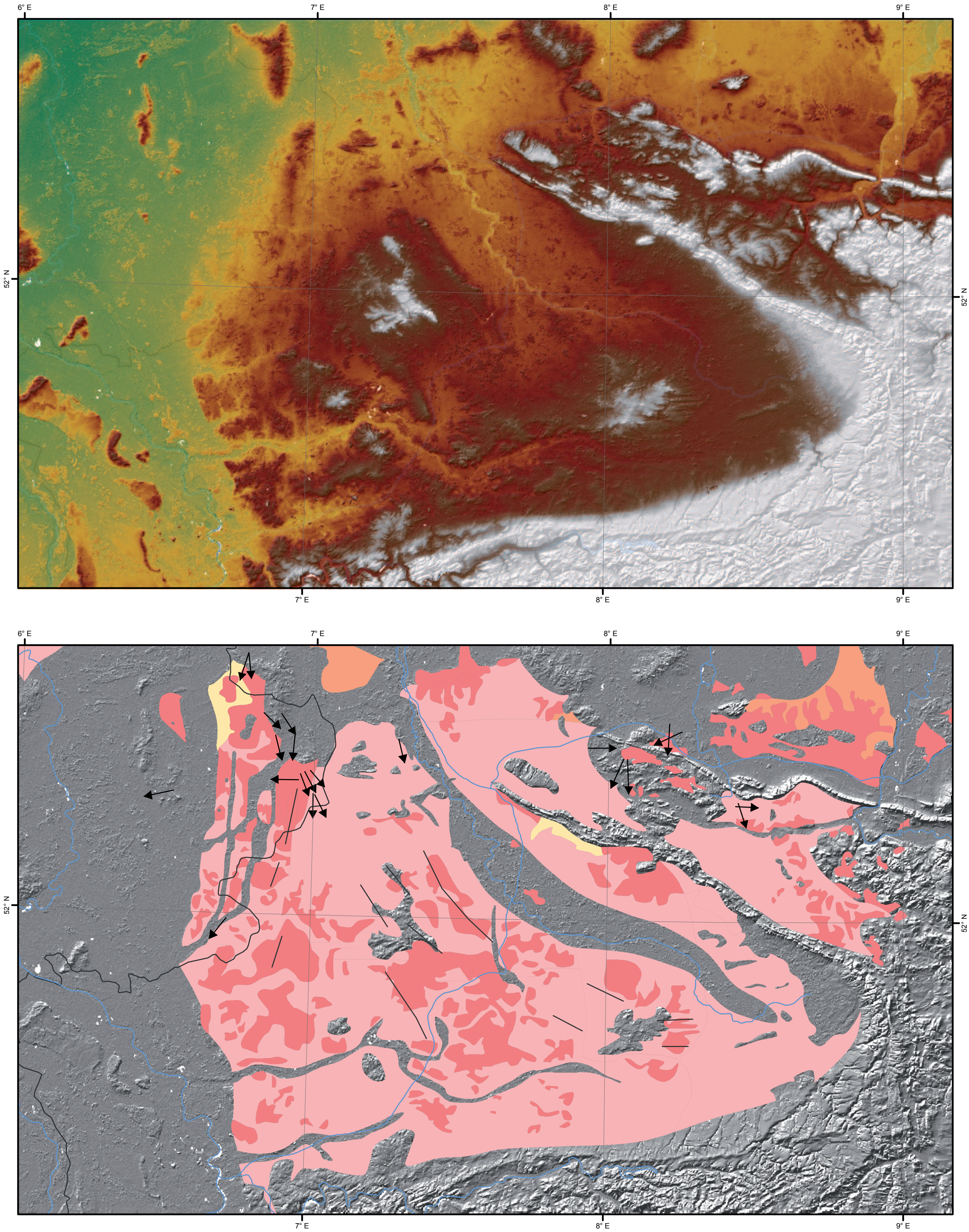


Figure 6.10
SRTM image of the till plateaus in Münsterland Enbayment. B) Deposits in the till plateaus with flutes deduced from the SRTM images. Furthermore till fabrics were plotted for the Weserbergland (after Skupin et al., 1993) and Twente (after: Rappol, 1993a and Ehlers, 1990). For the names of the till plateaus see figure 6.1

Till fabrics and lineations

Till fabric measurements of several authors (figure 6.6) were summarized in Ehlers & Stephan (1983) and Ehlers (1990a;b). On the Syke Geest and between Hamburg and Bremen the fabrics have a orientation between NNW-SSE and NNE-SSW. This is marked by fabrics from Bremen (Marczinski, 1968) and north of Bremerhaven (Meyer & Schneekloth, 1973). Most likely this flow direction can be coupled to the older Saalian till fabrics N-S in Hamburg (Ehlers 1990a;b). This flow direction was also found at the Hummling (Schröder, 1978), but this seems to be coupled to the Hondsrug ice flow. The NNW-SSE oriented flutes of the Syke Geest are truncated on the north by NE-SE running features. Therefore, this NNW flow must be older in this region than the NE-SW flow. On the eastern part of the Syke Geest diverting flutes are visible that truncate the flutes on the western side of the geest. This younger NEE-SWW flow (cf. SRTM and Ehlers & Stephan, 1983) may be correlated further to the east to the to NEE-SWW fabrics measured in the Elbe-Weser triangle by Lade (1980) and Marczinski (1968) and Stephan (1980). It continues to the west as a NEE-SWW and a NE-SW flow around the Hümmling (Schröder, 1978; Ehlers, 1990a). On the SRTM it can be seen that in fact both these two directions are also present Ostfriesland, where the NEE-SWW oriented ridges seem to be overprinted by the NE-SW structures. These pattern can be linked to the Drenthe till plateau ridges via till patches between these areas (e.g. Groningen).

6.1.7 Münsterland Embayment till plateau (Nordrhein-Westfalen)

In the Munsterbasin several till patches are present positioned on the Cretaceous chalk limestones(?). The thickness of the till ranges from several meters to several tens of meters (Speetzen, 1993; Klostermann, 1995). Residual tills with erratics are very common in this area, except for the Lower Rhine Embayment, here younger Rhine terraces dominate. Some tills were found that are interpreted as flow tills (Klostermann, 1995). Most of these erratics have a Scandinavian origin, but local erratics -derived from the Teutoburgerwald and the Bentheim region- are also abundant (Speetzen, 1993). In the southern part of the Munsterbasin almost no tills were found, whereas erratics are quite abundant (Skupin et al., 1993). Zandstra (1993) made a correlation between the Dutch tills and the tills found in Germany (figure 6.8). He found the Heerenveen group with mid-Swedish (Area II Fig..; Smaland, Dalarna) erratics. With a varying flint content. They distinguished an older part (clayey, Smaland erratics varying flint content) and a younger part (sandier, Dalarna erratics and high flint content) of which the former covers the largest area. Like in the Netherlands and Lower Saxony the red brown Assen till group with east Baltic components was found on top of the Heerenveen till group. The Assen till can be found between the 'eskers/ tunnel valley systems' in the Achterhoek and the Munsterland implying the distribution of this ice flow was limited between these two features. The Voorst and Rhenen till groups were absent in this area.

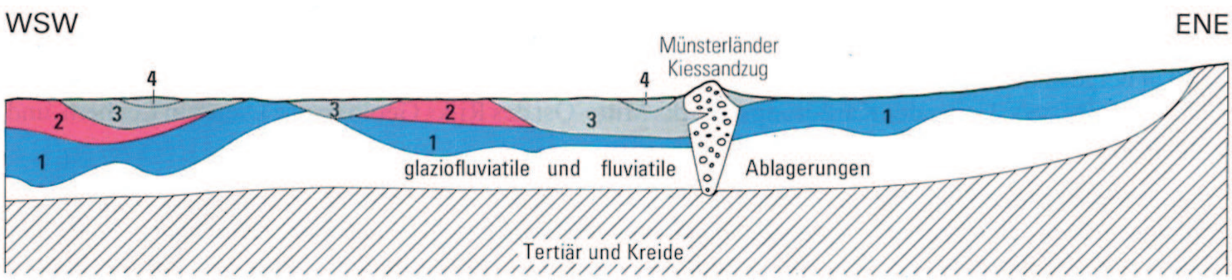


Figure 6.9
Schematical cross-section of the tills in the Westfälischen Bucht (from: Zandstra, 1993). Till 1 represents the Heerenveen till, (Smaland erratics, poor in flint); till 2 also represents the Heerenveen till (Dalarna, Smaland, rich in flint); till 3 and 4 represent the Assen till group containing eastern Baltic erratics, till 4 is the flint poor counterpart. The vertical scale is exaggerated.

Van der Wateren (2003) and Skupin et al, (1993; 2003) report drumlins on the southern edge of the Teutoburger Wald. These presumed drumlins indicate a NE-ward flow, probably as a consequence of severe diverging ice sheet that entered the Münster basin. They are hardly visible on the SRTM images. The topographical hights (e.g. eastern edge of the Beckumer Berge and north eastern side of the Baum Berge) show modification by the ice flow. Together with evidence from fabrics (figure 6.10) it was concluded by Skupin et al. (1993) that the ice flow was strongly diverting in the Münsterland Embayment Embayment. Based on this data Skupin et al. (1993) developed a glaciation phase model for Nordrhein-Westfalen (Chapter 7.2.1).
After the deglaciation and during the Weichselian large parts of this till was eroded on the western side by periglacial and deglaciation rivers. Between the till patches, erratics and sands occur as remnants. Therefore, it is very likely that the whole Münsterland Embayment Embayment was first totally covered by tills (Speetzen, 1993).

6.2 Ice-pushed ridges

In the research area different types of ice-pushed ridges are present. The pushed complexes range in size from very large to very small. Some of them were formed during an advance, others during a readvance or at the maximal stage. For their formative processes and content is referred to chapter 3 and chapter 5.3.

6.2.1 North Sea (Dutch part)

Features that were found 40 km west of Den Helder have been interpreted as ice-pushed ridges (Joon et al., 1990). These ridges are 10 meters high and have a similar orientation which is in line with the ice-pushed ridges on land. Laban (1995) found deformation structures west of the coast of Noord Holland on the edges of depressions that he interpreted as glacial basins. The exact location of the ice-pushed ridges was obtained from Busschers et al. (2008) and DGM (2009).

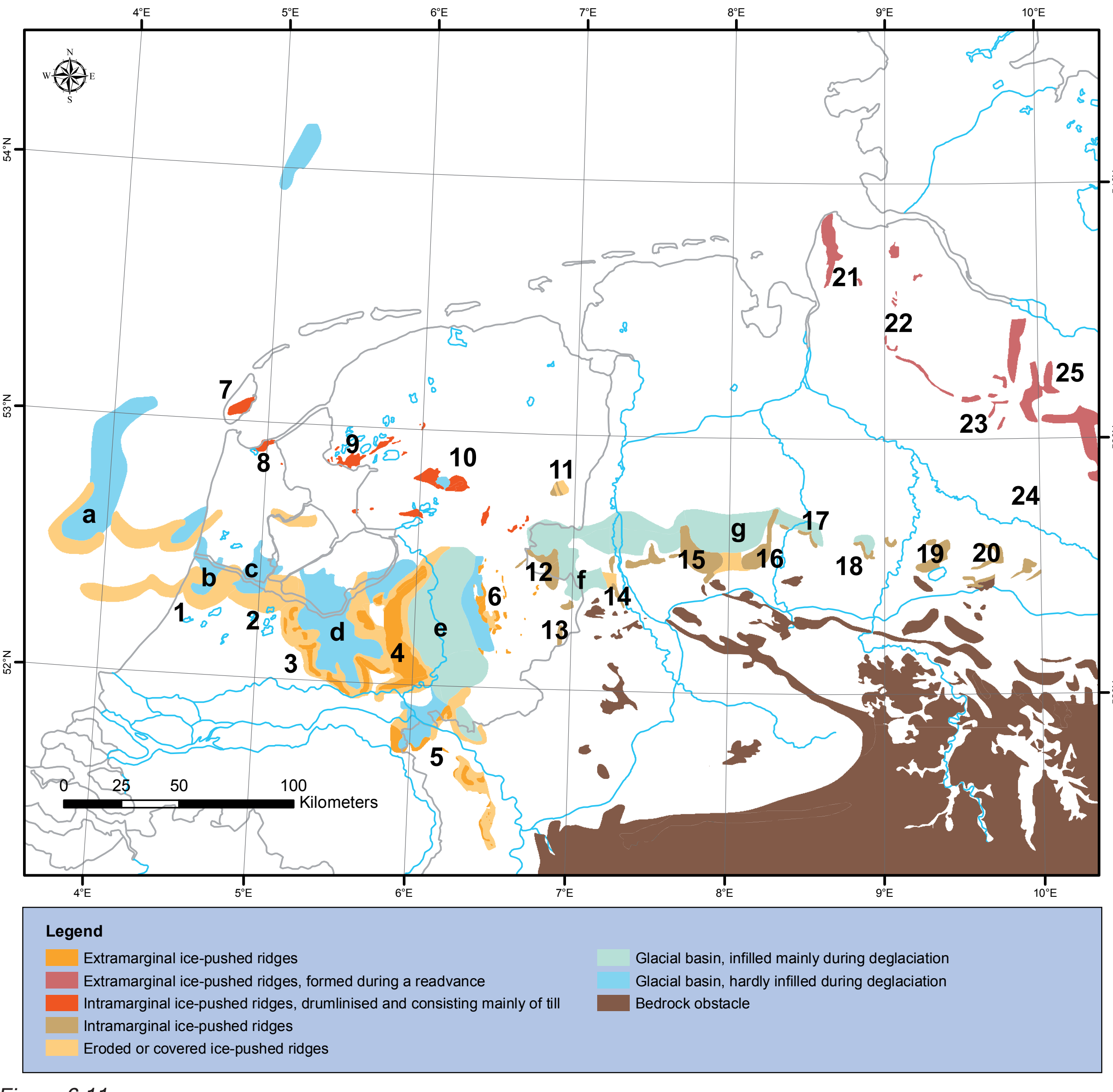


Figure 6.11
Ice-pushed ridges and glacial basins in the research area. Glacial basins: a = North Sea basin, b = Haarlem basin, c = Amsterdam basin, d = Gelderse Vallei, e = IJssel Valley, f= Nordhorn basin, g= Quackenbrück basin. ice-pushed ridges: 1= Haarlem, 2 = Amsterdam, 3 = Utrecht Ridge, 4 = Veluwe, 5 = Nijmegen-Kleve-Xanten ridges, 6 = Sallandse Heuvelrug, 7 = Texel, 8 = Wieringen, 9 = Gaasterland, 10 = Steenwijk ridges, 11 = Emmen, 12 = Itterbeck-Uelsen ridges, 13 = Oldenzaal-Enschede ridge, 14 = Emsbüren ice-pushed ridge, 15 = Fürstenauer or Ankumer Berge, 16 = Dammer Berge, 17 = Kellenberg, 18 = Böhrde, 19 = Schneerener Berge, 20 = Brelinger Berge, 21 = Altenward ice-pushed ridge, 22 = Lambeck ice-pushed ridge, 23 = Neuenkirchener ice-pushed ridge, 24 = Falkenberg, 25 = Garlstorfer ice-pushed ridge.

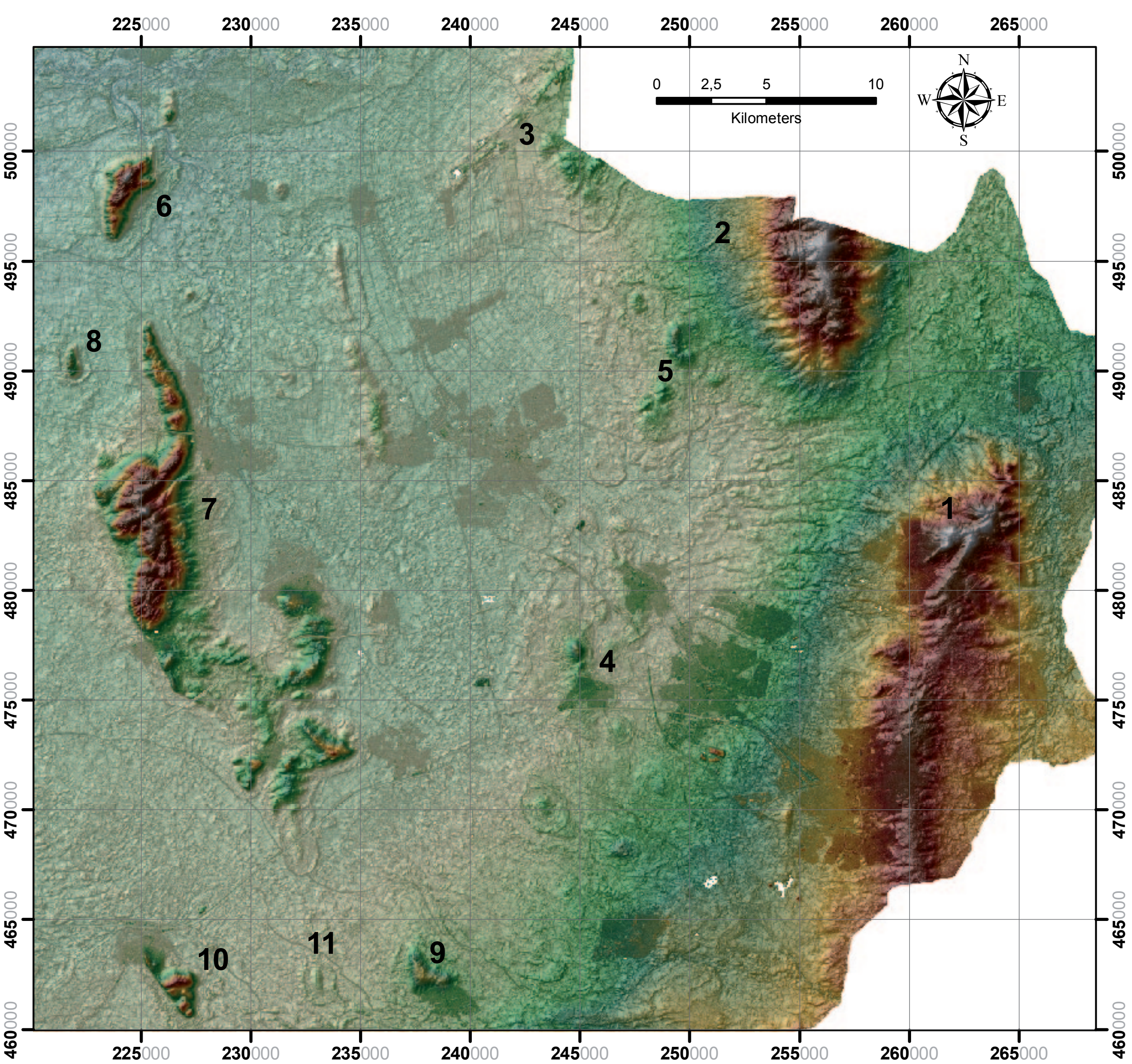


Figure 6.12
AHN image of the eastern Netherlands. 1 = Tankenberg (Enschede - Oldenzaal ridge), 2 = Ootmarsum ice-pushed ridge (southern extension of the Uelsen ice-pushed ridge), 3 = Sibculo ice-pushed ridge, 4 = Delden ice-pushed ridge, 5 = Albergen-Tubbergen ice-pushed ridges, 6 = Archemerberg and Lemelerberg, 7 = Sallandse Heuvelrug, 8 = Luttenberg, 9 = Needse Berg, 10 = Lochemse Berg, 11 = Geesteren ridge.

6.2.2 The eastern Netherlands

This area, situated south of the Drenthe plateau and east of the IJssel valley, has fine grained Tertiary sediments very near or at the surface with a thin Quaternary cover. The Itterbeck-Uelsen and the Enschede-Oldenzaal ridges are in line with the Rehburg ridges in Germany (Meyer, 1980; 1983). Like the ice-pushed ridges in the Rehburg line, most of the ice-pushed ridges in the eastern Netherlands have been overridden by the ice. Unlike the Rehburg ice-pushed ridges the eastern Netherlands ridges are typically shaped into craig-tail drumlins or flutes (figure 6.12) and contain a thick till cover (Van den Berg & Beets, 1987). Besides some smaller ice-pushed ridges of local interest are present in this region, which are shortly outlined below. They were probably formed by a lobated ice front from the north.

The Enschede-Oldenzaal ridge

This north-south trending ridge is about 9 km wide and 22 km long, its maximum height of 82m

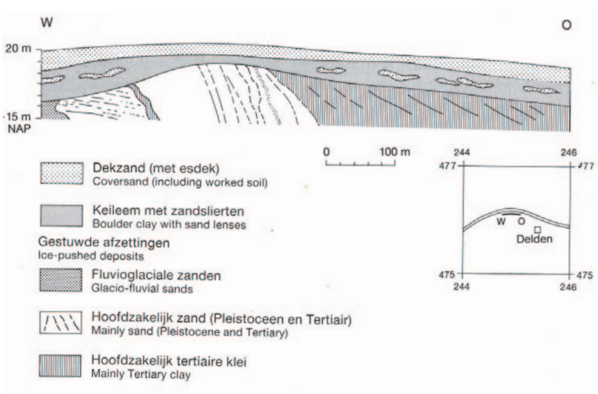
(Tankenberg) is located just north of Oldenzaal. Judging from the strikes and the morphology this ridge can be divided into two parts (Van den Berg & Den Otter, 1993; Kluiving et al., 1991): the northern part has a pronounced relief with tectonic structures trending west-east, coinciding with the Ootmarsum ridge. Whereas on the southern part has a more subdued morphology and N-S running strikes.

The northern part (north of the line Oldenzaal- De Lutte) consists mainly of large scale nappes of Tertiary sands and clays. Locally, thick tills can be found which are discussed in chapter 6.1.3. Thin nappes of Mid-Pleistocene sand and gravel deposits (Peize Formation) are also present. These nappes are mostly nearly horizontal and their fronts have a WNW-ESE orientation (Van den Berg & Den Otter, 1993). The ridge is considered to represent a segment of a ice-pushed ridge around the Nordhorn basin (Richter et al., 1951; Edelman & Maarleveld, 1958).

The southern part can be subdivided into two parts: the western part consists of pushed and replaced deposits. These deposits are mainly built up by Tertiary clays and some Mid-Pleistocene sands containing gravel (Urk Formation and Peize Formation). Presumably, these deposits were pushed from the area in between the Ootmarsum ridge and the Enschede- Oldenzaal ridge (Van den Berg & Den Otter, 1993). The eastern part of this ridge is built up of thick till (Voorst Till group- 20-40 m thick). The thickness of the till and the elevation of the ridge decreases towards the south (Van den Berg & Den Otter, 1993). After its formation the ridge was overridden by the ice from a NNW direction (Rappol, 1982; Kluiving et al., 1991).

The Sibculo ice-pushed ridge

Near Sibculo a small SW-NE ridge occurs, that is attached to the western edge of the Itterbeck-Uelsen ice-pushed ridge. Unlike the ice-pushed ridges around the Nordhorn basin, the Sibculo ice-pushed ridge only consists of sands of the Peize Formation (Van den Berg & Den Otter, 1993; Rappol, 1993a;b). According to Van den Berg & Den Otter (1993) the outcropping structures are part of a horizontal nappe. Rappol (1993a;b) states that this ice-pushed ridge consists of a large amount of thrusts dipping to the west, that have moved over a relatively small vertical distance ('bloktektoniek' figure 5.11) and that nappes may be present under the outcrops. According to Rappol (1993a;b) this ice-pushed ridge was formed by ice flowing from the east, then overridden by a ice flow from the NW (Rappol et al., 1991).



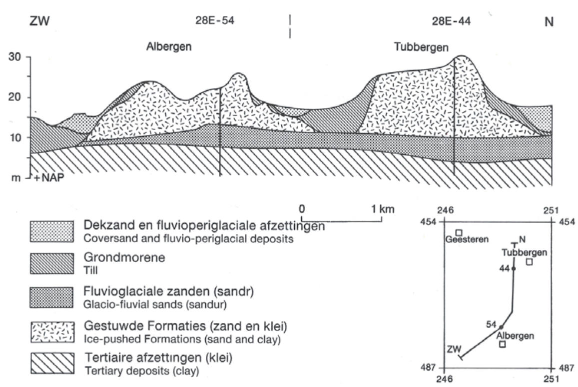
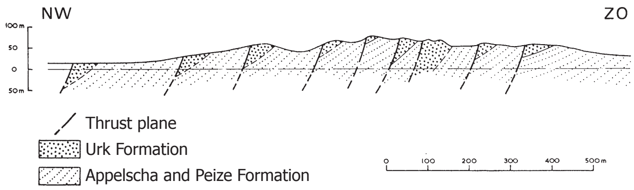


Figure 6.13 (left)
Schematic cross-section showing ice-pushed deposits unconformably overlain by till and cover sand, from Van den Berg & Den Otter, (1993).
Figure 6.14 (right)
Cross-section through the drumlin-shaped hills of Tubbergen and Albergen, from Van den Berg & Den Otter, (1993).



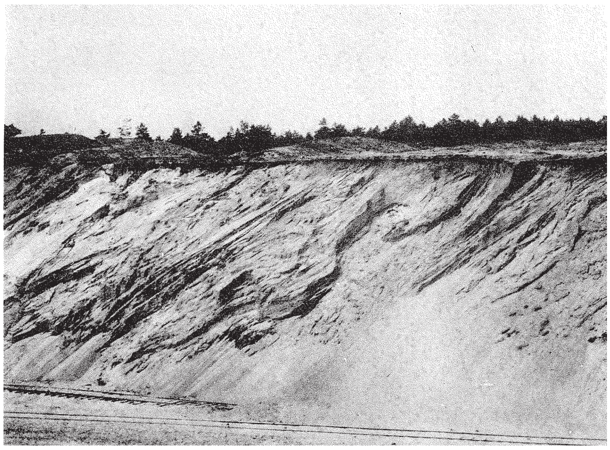


Figure 6.15 (above)
Photo of the northern section of the railway cut near Nijverdal showing the westerly dipping strata. From: De Jong, (1955); photo Burck 1922.

Figure 6.16 (right)
Schematic profile of the Archemerberg; After Rappol (1993) and De Jong (1962).

The Delden Tubbergen and Albergen ice-pushed ridges
The Delden ice-pushed ridge could be studied during the construction of a road (figure 6.13). The ice-pushed ridge consists of severely pushed and deformed marine Tertiary and the Peize Formation. Tertiary clays, that were most likely transported from 4 km E of the ice-pushed ridge, where this formation outcrops. The strike of the layers is N-S, dipping towards the east. Sometimes horizontally dipping layers were found. Besides, some pushed glaciofluvial deposits were found. The Delden ice-pushed ridge is covered with several meters till and on the northern side with fluvioglacial deposits of the Bruinehaar-Dinxperlo sandur (Van den Berg & Den Otter, 1993) zie h sandurs...
The Albergen and Tubbergen ice-pushed ridges have a height of maximal 30 m +NAP, and are elevated 15 m relative to the surrounding area. These ice-pushed ridges are located on a Tertiary substrate covered with sandur deposits (figure 6.14) and also consist of these deposits. The pushed deposits have been relocated by the ice from a direction SW on the undisturbed sandur. The Tertiary clays acted as a décollement for all these ice-pushed ridges (Van den Berg & Den Otter, 1993)

Sallandse Heuvelrug and Archemerberg
The Sallandse Heuvelrug and the Archemerberg are two ice-pushed ridges located at the eastern limit of the IJssel valley (figure 6.11). The Archemerberg is 5 km long and 2.5 km wide and has a maximal height of 80 m. The southern part of this ice-pushed ridge is called the Lemelerberg. This ice-pushed ridge is located several kilometers north of the Sallandse Heuvelrug.

The lithology of the ice-pushed ridges is quite similar the the lithology of the northern part of the Veluwe. It consist mainly of the Peize Formation, including the Hattem Bed. Also, Rhine deposits of the Urk Formation occur in the ice-pushed ridges (De Jong, 1955; Maarleveld, 1956).

In the Archemerberg, thrust sheets occur as thick as 100 m (De Jong, 1962). The strike of the thrusts generally runs parallel to the orientation of the ice-pushed ridges. The layers dip 25-45o to the N in the north eastern part of the moraine, south of the highest point the dip is N (De Jong, 1952; 1955). Near Nijverdal a railway section perpendicular to the N-S strike of the thrusts provided a cross section in the northern part of the Sallandse Heuvelrug (figure 6.15 - De Jong (1955; 1962) described the observed structures as 'wedges with a thickness of about 5 to 20 m dipping on average 40o to the west'. The thickness was thought to be due to the thickness of the permafrost, but later authors demonstrated that this is not necessarily a prerequisite.
Like on the Veluwe a connection exists between these different outcropping morphologies and the geomorphology on both ice-pushed ridges (figure 6.16 - De Jong, 1955). The gravelly Urk Formation was slightly more resistant to post-Saalian erosion and forms the core of the small superimposed

ridges that run parallel to the striking of the ice-pushed ridge. This can also be seen on the figure 6.11. It can also be seen that the northern part of the Sallandse Heuvelrug clearly differs from the southern part. The northern part is also connected to the small Luttenberg in the NW, forming a horse shoe shaped ice-pushed ridge which was most likely formed by an ice lobe from the N (Rappol et al., 1991). The southern part of the Sallandse Heuvelrug was most likely formed by a combination of ice pressure from the E and W.
The orientation of the thrusts in the Archemerberg indicates that this ice-pushed ridge was formed from a W and NW direction. De Jong (1955) states that the NE-SW ridges cut off the EW ridges, indicating a first push event from the north (Vecht region), and subsequently from the IJssel valley.
The Archermerberg and Sallandse Heuvelrug were probably connected during the glaciation. They were then separated by large meltwater streams.

Needse Berg, Geesteren ridge and Lochemse Berg
The Needse Berg is a small isolated ice-pushed ridge in the Achterhoek region in the Eastern Netherlands. It is located several kilometers east of the Lochemse Berg. The ice-pushed ridge has a peculiar boomerang-like shape (figure 6.12). The main part of the ice-pushed ridge consists of Tertiary marine deposits, the Breda Formation. Also, the coarse Rhine sands (Urk Formation) are present at the western flank. Within these sands a fluvial clay layer from the Holsteinian is present, known as the 'Neede Clay'. The strata in the Needse Berg are clearly disturbed by the pressure of the ice, although large thrusts or nappes were not found (Pierik et al., 2010). The pushing direction was most likely from the NNE to NE.
The Lochemse Berg is a small isolated ice-pushed ridge in the Achterhoek region in the Eastern Netherlands. It is located several kilometers west of the Needse Berg and its highest point is 50 m asl. It is mainly composed of the Peize and Appelscha Formation. Also the Urk Formation is included and like in the Needse Berg, also clayey Holsteinian Rhine sediments are present. Presumably, this ice-pushed ridge was formed by ice pushing form the E to NE (Nijhof, 1993; Van den Berg et al., 2000). In the southwestern side the ice-pushed ridge is sharply eroded (figure 6.12), probably this happened by the Rhine during the deglaciation.
In between these ice-pushed ridges, near Geesteren, a small N-S trending ice-pushed ridge is present covered with till. This may be interpreted as a drumlin (Van den Berg et al., 2000).

6.2.3 The central Netherlands

The Utrecht Ridge
The Utrecht Ridge (Utrechtse Heuvelrug) is located on the southwestern edge of the Gelderse Vallei and it is a series of ridges composed of imbricated thrusts sheets (Aber et al., 1989; Ruegg & Burger, 1999). It rises 50 m above sea level and has an asymmetric slope profile. The ridge is composed of several minor ridges (figure 6.18). In the northern part (Gooi) several small ice-pushed ridges occur (Rappol et al., 1994). The Amersfoort ridge has EW running strikes and an EW orientation, which is perpendicular to the Utrecht ridge, indicating a younger frontal pushing event (cf. Maarleveld, 1981). Van Balen (2006) and Busschers et al. (2008) described a section located on the edge of both ridges in which both pushing form the N and the E could be seen. The eastern side, which is adjacent to the glacial basin of the Gelderse Vallei is much steeper (5-15 degrees) that the western side (2-5 degrees) where sandurs are present (chapter 6.4.2).
The Utrecht Ridge is composed of the Waalre, Urk and Drente Formation (Van der Wateren, 1985; Ruegg & Burger, 1999). The Waalre Formation contains fine sand of southern provenance (Rhine catchment), loam and clay that served as a décollement. This ridge developed on the NE rim of the relative upthrown fault block of the Peel high (Van der Wateren, 1995). The Urk Formation contains coarse sand and gravel from the river Rhine predominantly, which are considered to be suitable for the

formation of the large ice-pushed ridge (Van den Berg & Beets, 1987).

The Utrecht ridge was formed by an ice lobe that entered the Gelderse Vallei (e.g. Aber et al., 1989). The pushing of the ridge took place along the marginal side of the lobe (Van der Wateren, 1981; 1985). It resulted in imbricated thrust blocks striking parallel to the ridge and dipping on average 35 to 40 degrees NNE. The strikes can be seen in figure 6.18. 20 nappes were found and the thickness varies from several meters to 25 meters. In the northern Gooi part two series of ice pushed ridges with imbricate structures dipping towards the east are present. They indicate two pushing events from the east (Rappol et al., 1994; Ruegg & Koopman, 2010). The small isolated ice-pushed ridges of Soest and Baarn indicate that severe erosion took place in northern part of the Utrecht Ridge, during or after its formation. The formation of the Utrecht ridge occurred between 168 ka and 150 ka, these ages are derived from respectively the youngest Urk Formation deposits and the sandur deposits (Van Balen et al., 2007; Busschers et al., 2008).

Tills or deformation structures on top of the hill are rarely found. However, some erratics were found on top of the ice-pushed ridges. Van Balen (2006) found a sandy till remnant near Soesterberg indicating slight overriding. Like most ice-pushed ridges, the Utrecht ridge was already subjected to severe erosion during the melting of the ice in the Saalian. Large meltwater valleys were formed, of which the Darthuizer Poort is the largest and most well known (Berendsen & Bijnen, 1973). In the Gooi area, the pattern of small ridges suggest significant meltwater erosion. Here, the Utrecht ridge was also eroded by lateral marine erosion in the Eemian. In the Wechselian periglacial erosion occurred, mainly due to surface snow melt under permafrost conditions forming snow meltwater valleys.

The Veluwe

The Veluwe is the largest and most striking ice-pushed ridge complex of the research area. It is composed of several ice-pushed ridges, the highest point of 109 m is located in the very southeast. In the northern part the ice-pushed ridges are flanked by kame terraces (chapter 6.6), on the southern part a sandur can be found, known as the Schaarsbergen sandur (figure 6.R). The strikes of these ice-pushed ridges were mapped by Maarleveld (1953, 1981). Based on this mapping he made a phase model for the glaciation of this area (outlined in chapter 7.1.) Nowadays, this research can be updated by using the high resolution elevation models (figure 6.19).

The largest ice-pushed ridge is the Eastern Veluwe ice-pushed ridge running from Hattem to Arnhem. The strikes run more or less parallel to the ice-pushed ridge (figure 6.18). Around the Woldberg the strikes have a more peculiar orientation around the Woldberg. They

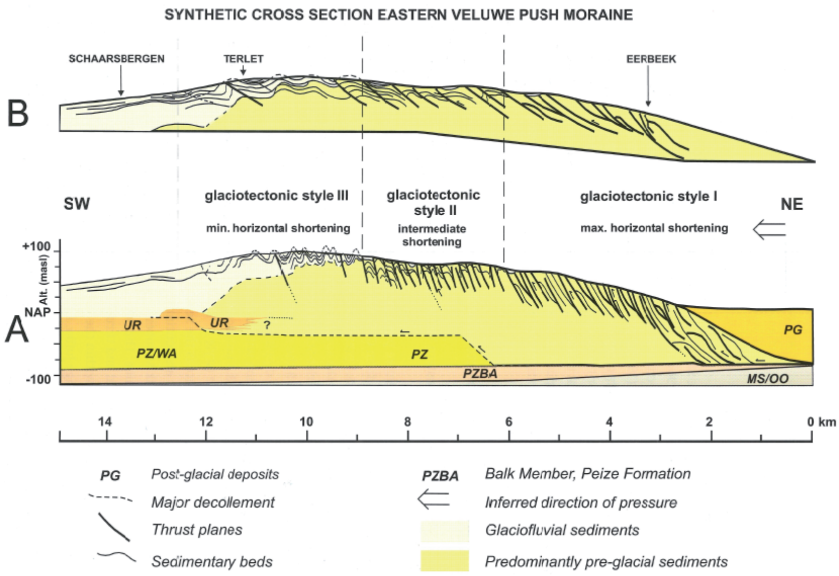


Figure 6.17
Synthetic model of the southern part of the eastern Veluwe ice-pushed ridge (for location see figure 6.18). The bold lines indicate thrusts and the thin lines indicate the deformed sedimentary strata. In section A the internal structures are at scale and in panel B the size of the structural elements is not at scale, but drawn according to true angles of dipping features. The Balk Member (Peize Formation) acted as principle décollement. From: Bakker (2006).

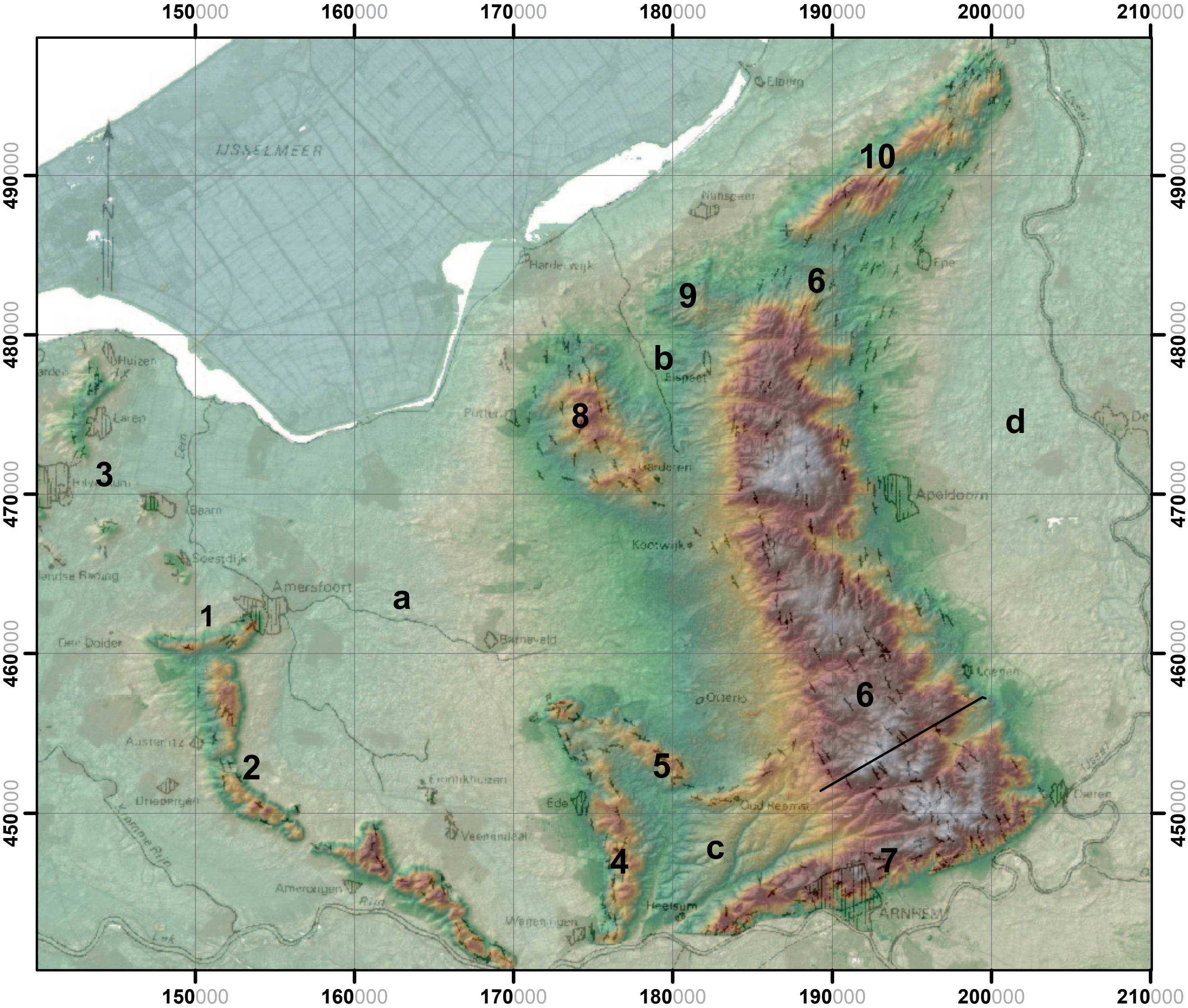


Figure 6.18
AHN high resolution elevation image of the Veluwe and Utrecht ridge. The strikes from Maarleveld (1981) are plotted as well. These strikes tend to dip towards the direction the ice came from. On the distal parts the dips can differ because folding instead of thrusted strata were measured (compare crossveluwe). Especially in the northern part and at the Stakenberg the geomorphology coincides with the internal structure of the ice-pushed ridge. Note the formation of substantial ice and snow water valleys formed respectively during the deglaciation and the periglacial conditions of the Weichselian. 1 = Amersfoort ridge, 2 = Utrecht Ridge, 3 = 't Gooi, 4 = Ede ridge, 5 = Oud-Reemst ridge, 6 = Eastern Veluwe ice-pushed ridge, 7 = Arnhem ice-pushed ridge, 8 = Garderen ridge, 9 = Stakenberg, 10 = Woldberg. a = Gelderse Vallei, b = Leuvenemsche Beek kame, c = Schaarsbergen Sandur, d = IJssel Basin. The cross section of figure 6.17 is indicated in the south-eastern part of the Veluwe.

seem to be displaced by a pushing event from the NW that formed the Woldberg. Some quarries in the northern part of this ice-pushed ridge -near Wapenveld and Hattem– were studied by Zandstra (1971). The sediments in the ice-pushed ridge comprise the entire middle Pleistocene sandy fluvial succession. The base of the thrusts consist of clays of Tiglian age (Balk Member, Peize Formation, Early Pleistocene), which acted as a décollement in the northern part of the IJssel basin (Zandstra, 1971; Van den Berg & Beets, 1987) as well as in the southern part of the ice-pushed ridge (Bakker, 2006).

Using GPR (chapter 2.2) Bakker (2006) revealed that imbricated thrusts form the internal structure of the southern part of this ice-pushed ridge. Zandstra (1971) also found 40-60° ESE dipping thrust sheets, with a maximal thickness of 88 m. This structure is confirmed by the clear strike parallel ridges from Apeldoorn to Hattem, visible on the AHN (figure 6.18), which are unlikely to resemble sub-horizontal nappes (Bakker, 2006). In the more distal parts (i.e. more to the west) folds are present, and in the most distal parts the folds dominate (figure 6.17). In the northern part of this ice-pushed ridge thrusting is also the most likely structure.

Bakker (2006) concluded, based on sandur deposits configuration, that the Arnhem ice-pushed ridge was formed at the same time as the East Veluwe ice-pushed ridge, or probably after the eastern Veluwe ice-pushed ridge. In contrast to Maarleveld (1953; 1981), Bakker (2006) suggests that the eastern Veluwe ice-pushed ridge was formed during one single pushing phase which caused large-scale imbrication of displaced units. The Stakenberg, south of Nunspeet, consists of pushed glaciofluvial (kame terrace) deposits (Maarleveld, 1981). Its position relative to the Woldberg indicates same age (Maarleveld 1981).

The current morphology of the Veluwe was severely affected by erosion and weathering since their formation, starting in late Saalian times and continuing in the last glacial. This is evident from their plateau like geomorphology (Ten Cate & Maarleveld, 1977), the meltwater channels and from the fill of the Amersfoort and IJssel glacial basins Busschers et al. (2008). The lithology, internal structure and location and orientation of the thrust sheets co-determined the resultant geomorphology (Teunissen, 1961; Bakker, 2006).

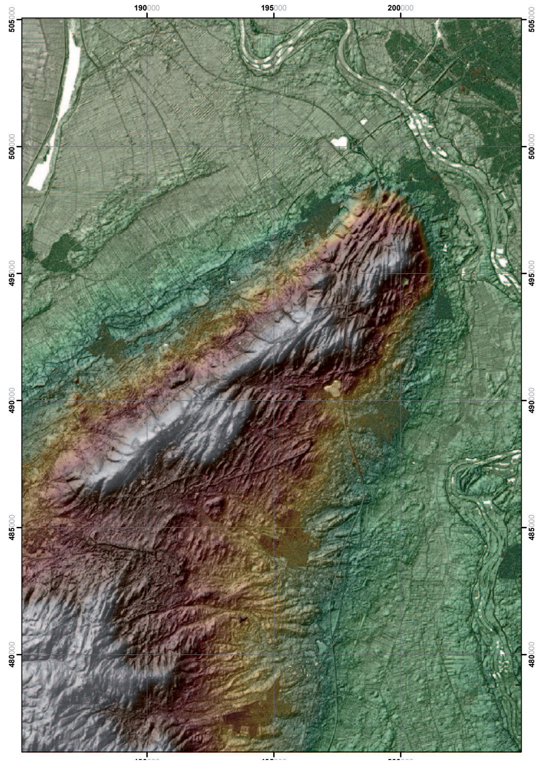


Figure 6.19
Detail of the AHN image of figure 6.18. The coincidence of the geomorphology and the internal structure can be seen very clearly on this part of the Veluwe. The striking of the thrust planes seem to be pushed away by the formation of the Woldberg.

6.2.4 Northern Netherlands

The southern edge of the Drenthe Till Plateau is marked by a series of ridges on the line Texel-Wieringen-Gaasterland-Steenwijk-Hoogeveen-Coevorden (figure 6.11). These ridges have several features in common (Van den Berg & Beets, 1987):

a. They are small compared to the moraines in the central parts of the Netherlands.

b. Only the uppermost 10 to 20 meters of the pre-glacial sequence are involved in tectonic dislocation, and consequently the glacial basins found with these ridges are shallow or even absent.

c. They are often associated with the Voorst Till (chapter 6.1). This till is rare in the research area except for these ice-pushed ridges. The till in these ridges is quite thick (>10m) and has been imbricated with preglacial sediments (Ter Wee, 1962).

d. Fluvioglacial sediments such as ice-contact fans or meltwater deposits are not involved in pushing.

e. The ridges have been fluted and drumlinized in a NE-SW direction (Brouwer, 1950; Zonneveld, 1975) and consequently, have been overrun by the ice-sheet.

These ice-pushed ridges clearly differ from other ice-pushed ridges in the research area because they are relatively small and they contain tills.

The ridges on the line Texel-Hoogeveen are located at the margin of the eastern branch of the Holsteinian Rhine (Van den Berg & Beets, 1987). The décollement was formed west of Hoogeveen by Holsteinian marine clays and east of Hoogeveen by Tertiary and Mesozoic clays (Van der Wateren 1995; Van den Berg & Beets, 1987).

Whether these ridges were formed during a long stagnation of the ice front or during a continuous ice advance is still an issue of debate. Rappol & Kluiving (1992) consider that these ridges were ablation moraines that were pushed after their formation. As the formation of the ablation moraine takes some

time, they would indicate that the ice front stagnated for some time before it reached the maximal extent. Van der Wateren (2003) considers these ridges to be stacked till sheets (style Ec, chapter 5.3).
Another issue of debate is the correlation with the Rehburg line in Germany (chapter 6.2.5). Thome (1959) and Zonneveld (1975) correlate these ridges to the large ice-pushed ridges in Twente and the German Rehburg line. Ter Wee (1962), Jelgersma & Breeuwer (1975), Van den Berg & Beets (1987) correlate the Rehburg phase to the ice-pushed ridge in central Netherlands. This is however very difficult to state because the Rehburg line does not necessarily indicate one single time line (Van der Wateren, 1995).

Some ice-pushed ridges of the line Texel-Gaasterland-Steenwijk-Coevorden have been drumlinized and fluted in a NE-SW direction (Brouwer, 1950; Zonneveld, 1975; Van den Berg & Beets, 1987). This can be seen very nicely in the AHN images (figure 6.2). In the eastern Netherlands features were found that were interpreted as crag and tail drumlins. e.g. Enschede-Oldenzaal ridge (Van den Berg & Den Otter, 1993)
On Wieringen and Texel ridges occur that also consist of thick pushed till deposits and preglacial deposits (De Gans, 1994). In the till on Wieringen a NW-SE fabric was found (Rappol 1991b wieringen).
In the area Winschoten-Scheemda features occur that have been interpreted as drumlins (Edelman & Maarleveld, 1958) or ice-pushed ridges (Ter Wee, 1962). Lineations on the features show a NE-SW ice flow. They consist of Elsterian deposits ('Potklei') and are covered with the Heerenveen till (Speetzen & Zandstra, 2009). Because of their little elevation and position between deglaciation channels and Holocene marine estuarine channels it is more likely that these features form the last remnants of a till plateau that was present here.

Steenwijk ridges
Around Steenwijk several ridges occur: The Woldberg (not to be confused with the Woldberg in the northern Veluwe), the Steenwijkerkamp and the Bisschopsberg.
The main part of the ridges is composed of tills (Voorst till Group) which varies from 2 to 5 meters thick, but it can also be 20 m thick (Ter Wee, 1966). Early Saalian periglacial deposits (Boxtel Formation, Drachten Member) are also present in the ridges, in elevated pushed position up to 25 m asl (Ter Wee, 1966; Rappol & Kluiving, 1992; Koster, 2010). Morphologically, the ridges are formed of NE-SW trending parallel striking smaller ridges (figure 6.2). They can be interpreted as drumlinized or fluted ridges which formed when the hills were overridden.
Ter Wee (1966; 1983) described a adjacent small basin north at 15 m below sea level, which he interpreted as a glacial basin were the ice lobe that formed the ridges was located.
According to Rappol & Kluiving (1992) the sequence of events near Steenwijk was as follows: The ice front stagnated at this edge forming an ablation moraine (Voorst till group). This moraine had a horseshoe shaped form (Rappol & Kluiving, 1992; Koster, 2010). When the ice prograded again, pushing occurred and the till was deformed and subsequently overridden, drumlinizing the ridges and depositing the Heerenveen tills on

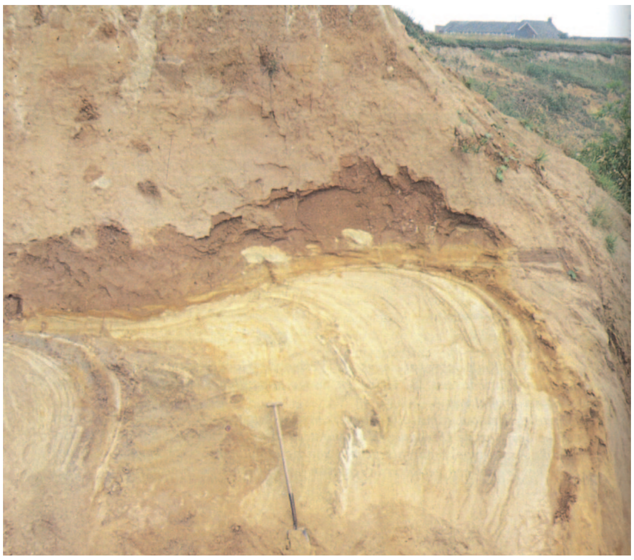


Figure 6.20
The upper section of the ice-pushed deposits is clearly deformed by the ice flow from the right, on top these deposits till was deposited. The photo was taken in the Ten Boer pit near Emmen (from: Rappol & Kluiving, 1992).

top of them (Koster, 2010). A similar sequence was also found near De Lutte (chapter 6.1). During the deglaciation and in the Weichselian the Rhine eroded the southern parts of the complex.

Emmen

Directly north of the village Emmen a glaciotectonic structure is present, which differs from the other ridges in the northern Netherlands. The ridge cannot be found in the present topography as it has been overprinted by the NNW-SSE running Hondsrug structures (Ruegg & Zandstra, 1977). Nappes or thrust structures exposed in the quarry Ten Boer have a thickness of over 40 m and consist of Pliocene and lower and middle Pleistocene deposits (Ruegg & Zandstra, 1977; Ter Wee, 1979; Van den Berg & Beets, 1987). On top of these structures tills were deposited (figure 6.20). The orientation of the thrusts indicate pushing from the north. The adjacent glacial basin has never been found, probably it is hidden below the fluvioglacial Hunze valley, east of the Hondsrug (Van den Berg & Beets, 1987)

6.2.5 The Rehburg line

This Rehburg line consists of a line of ice-pushed ridges and can be traced from Hannover into the Netherlands and the North Sea (figure 6.6). Initially, these moraines were considered to be recessional moraines. Nowadays, these ice-pushed ridges are considered to be formed at an early stage in the glaciation, after which they were overridden by the ice (Meyer, 1980; Van der Wateren, 1995). Although these ice-pushed ridges have been overridden, they do not seem to be as clearly fluted or drumlinized as happened in the Northern Netherlands (figure 6.2 and 6.6).

Besides, this line was traditionally attributed to one stagnation phase: the 'Rehburg Phase' (Woldtstedt, 1928; Lüttig, 1958). Van der Wateren (1995, 2003) concludes that the 'Rehburg Phase' as it was originally defined does not exist. The Rehburg line merely consists of a series of ice-pushed ridges rimming the northern marges of the highlands, which were formed during the Drenthe advance, but not necessarily synchronously. Not glaciological conditions, but the substratum is the main control in the formation of the Rehburg line line (chapter 5.2; figure 5.9). The substratum consists of Mesozoic and Tertiary clays that could serve as a décollement.

How the Reburg line continues in the Netherlands is still issue of debate. Van den Berg & Beets (1987) and Van der Wateren (1995) correlate the Rehburg ice-pushed ridges to the ice-pushed ridges in the central Netherlands. Rappol (1991a) correlated them to the ice-pushed ridges in the Northern Netherlands.

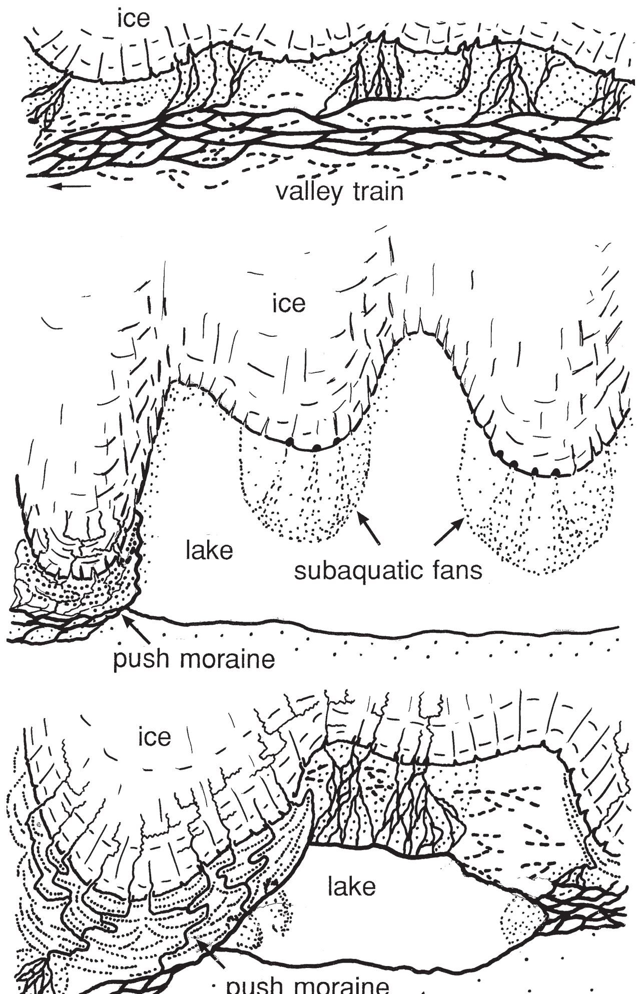


Figure 6.21
The Dammer Berge and Fuerstenauer Berge probably formed from glaciofluvial sediments from ice marginal rivers (Weser) draining towards the west. Due to the prograding ice front the water was trapped and a lake formed in which glaciodeltaic sediments were deposited. At a certain moment, the condition favoured pushing and the ice-pushed ridges were formed from these sediments (from: Van der Wateren, 1995).

The Dammer and the Fuerstenauer Berge (Rehburg line)

These ice-pushed ridges form a semi-circular series of ridges 40 kilometers across and between 40 and 145 meters above sea level. They are built up of preglacial material (see chapter 3) in the region: Tertiary sediments and some Cretaceous sediments, also white sands that correlate to the Appelscha Formation in the Netherlands were found. Also, sediments from the river Weser occur, the Oberterrasse and Mittelterrasse are incorporated. The latter contains a lot of pink sands from the Red Buntsandstein. The most dominant lithology are the glaciofluvial sediments deposited at the ice margin just prior to, during and subsequent to the formation of the push structures (i.e. pre-, syn- and post-tectonic glaciofluvial sediments, respectively). A major part of these sediments appears to be made up of subaquatic sediments with the characteristics of proglacial deltas and subaquatic fans which formed during the progradation of the ice front (figure 6.21 - Van der Wateren, 1994; 1995, 2003).

The tectonic structure (figure 5.12) consists mainly of sub-horizontal nappes which have been displaced horizontally from the Quakenbrück basin, in some cases as far as 6 km. Six subhorizontal nappes have been found with an aspect ratio of 20:1 to 50:1 (Van der Wateren, 2003). Each nappe is normally composed of coarse grained glaciofluvial or Tertiary sands and gravels that have been displaced over a substratum of ductile clay or silt. Folding tectonical structures form a minor part. The lack of thick sandur plain deposits makes it unlikely that the ice stagnated for a long time. From the relation between the glaciofluvial deposits and the development of glaciotectonic structures Van der Wateren (1995) concluded that the overriding of the ice-pushed ridge occurred relatively soon after its formation, several years or decades.

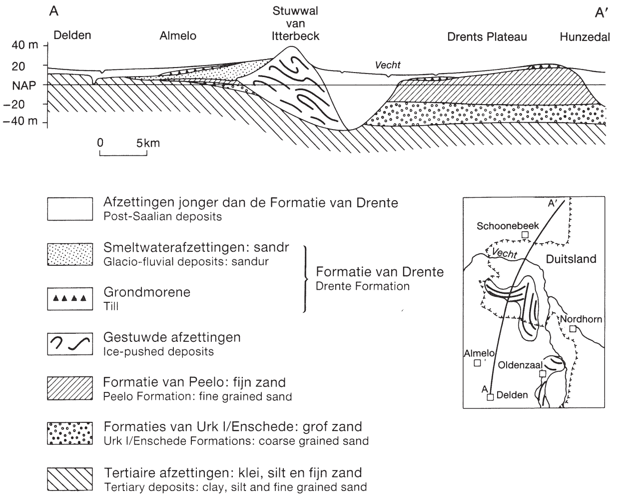


Figure 6.22
Cross-section from the Drenthe plateau to Twente showing the position of the Itterbeck ice-pushed ridge and its glacial basin in relation to the hydrogeological base of the Tertiary deposits (from Van den Berg & Den Otter, 1993).

Itterbeck-Uelsen ridges

The two Saalian ice-pushed ridges of ltterbeck - Uelsen, formed during the Older Saalian glaciation, and are part of the German Rehburg line (Kluiving, 1994). The Uelsen ridge trends N-S, is about 5 km wide and it has a maximum elevation of 90 m +SL. The ridge is flanked by the Nordhorn glacial basin. The Itterbeck ridge is a W-E trending curved ridge of 2 a 3 km wide and up to 70 m high. It flanks the Wilsum glacial basin (50-75 m -SL).

The lithology of the ice-pushed ridges consists of Tertiary and Quaternary deposits (figure 6.23 - Kluiving 1994). The ice-pushed ridges were formed where the Tertiary clays occur

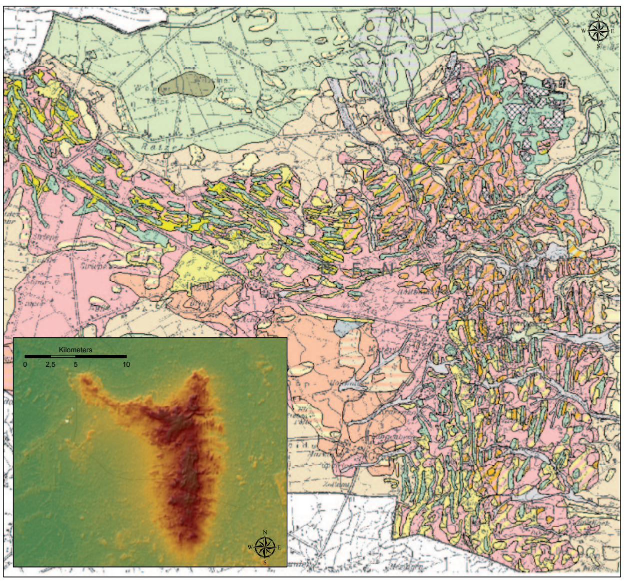


Figure 6.23
SRTM image and geological map of the Itterbeck-Uelsen ridges. The maps do not have the same scale. Both on the SRTM image and the geological map the thrust sheets at the surface can be seen. Colours on the geological map: yellow:Tertiary?; Green: Middle Pleistocene deposits; Light pink: fluvioglacial deposits; pink: tills (from: geological map 1:50.000 - LBEG, 2006).

closer to the surface were they could easily act as a décollement (figure 6.22 - Van den Berg & Beets, 1987). The Early Pleistocene sands consist of coarse fluvial sands from eastern provenance (Peize Formation and the Appelscha Formation). Locally, glaciofluvial deposits are prominent as a result of reworking of older deposits by meltwater. They have been deposited during the formation of the ice-pushed ridges (Kluiving, 1994). On top of the ice-pushed ridges patches of tills and deformation structures were found, indicating overriding. The southern tip of the Uelsen-Oldenzaal ridge, south of the line Ootmarsum-Reutum consists only of till (Van den Berg & Den Otter, 1993).

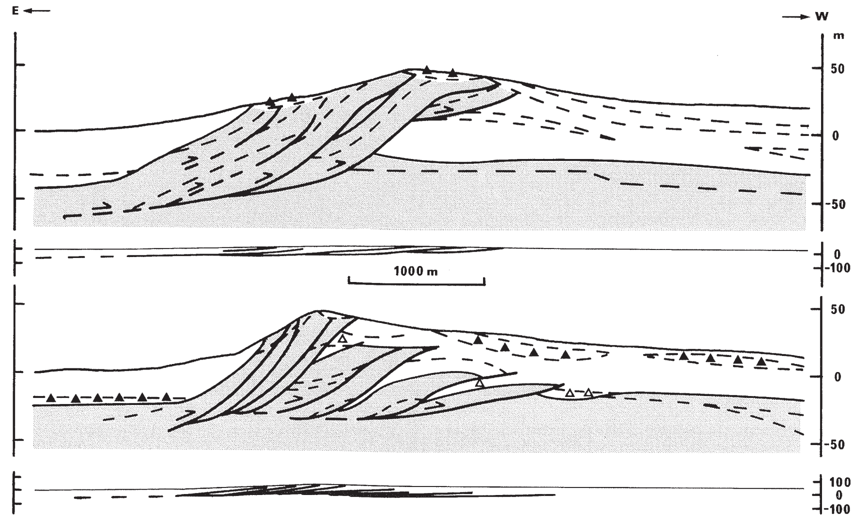


Figure 6.24
Internal structure of the Lamstedt ice-pushed ridge (from Van der Wateren, 1995; after Van Gijssel, 1987).

The internal structure of the ridges was first described by Richter et al. (1951). They concluded that the ridges were built up of thrust sheets. Kluiving (1994) demonstrated that this approach is too simple and that the dominant structures are large, shallow and flat-lying glaciotectonic nappes that have been displaced over distances of at lest 1 km in the direction of tectonic transport. The strikes of the thrusts in the Dutch part of the Uelsen moraine are N-S in the northern part and NW-SE in the southern part of the ice-pushed ridge. The thrust mostly dip towards the E and N indicating that the pushing direction was from the NE, which means that the ice-pushed ridge was built up from material from the Nordhorn Basin (Van den Berg & Den Otter, 1993). A relation between the internal structure and current morphology is locally present (figure 6.23). Overprinting relations of two deformation phases gave a relative age difference between the two ice-pushed ridges. Kluiving (1994) concluded that the Wilsum ice lobe (N) advanced after the Nordhorn ice lobe (E).

6.2.6 Northern Lower-Saxony

The Lamstedt Moraine and Altenward Moraine

These ice-pushed ridges were formed in a smooth subglacial erosion landscape with a thin cover of tills and glaciofluvial beds, overlaying Tertiary sediments. Both the Lamstedt ridge and the Altenwald ridge run parallel to a series of N-S trending outcropping salt domes (Kuster & Meyer, 1979). These salt domes favoured the conditions for the formation of ice-pushed ridges by pushing the Tertiary clays close to the surface (Kuster & Meyer, 1979 – figure 5.9 en figure 3.3)

According to Höfle & Lade (1983) the style of deformation is characterized by thrust faults and the tectonic transport was to the west causing eastward dipping strata. These thrust faults were either pushed frontally or laterally and dip 2 -8 degrees. Van Gijssel (1987) concluded based on more evidence that the ice-pushed ridge was made of nappes (figure 6.24). These nappes moved on a décollement of Tertiary clay which had been pushed up by the salt domes. The nappes consist of Holsteinian clay, silt from the Elsterian tunnel valleys, Drenthe-1 tills and glaciofluvial deposits (Höfle & Lade, 1983; Van Gijssel, 1987).

This ice-pushed ridge was probably formed during a re-advance of the retreating ice, this is concluded from the older tills that are present in this moraine (Höfle & Lade, 1983 - figure 5.15). The moraine

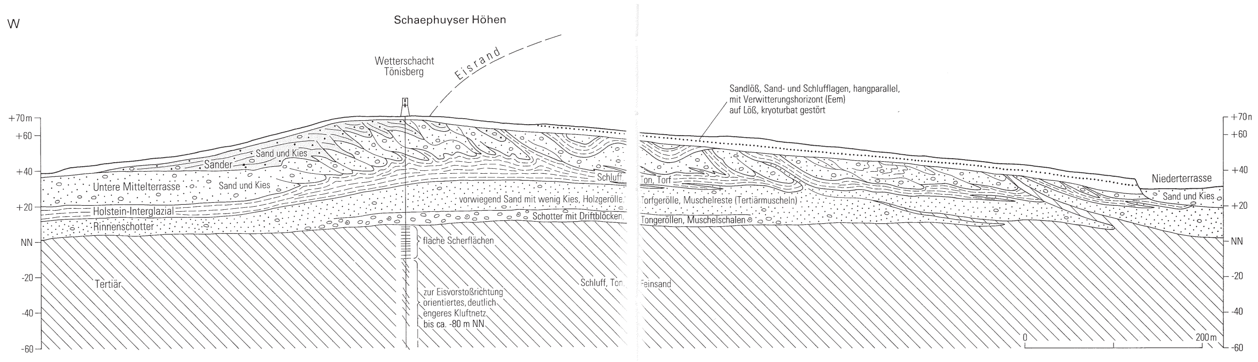


Figure 6.25
Internal structure and composition of the Schaephuyser Hohen, for explanation see text (from: Thome, 1984a). The ice-pushed ridge is the southern most elongated ridge in figure 6.11

seemed to be at least partially overridden because deformation structures and lodgement tills were found on top of the moraine. At the distal side of the moraine the Drente II till was found which suggests a formation of the ice-pushed ridge during a re-advance (Höfle & Lade, 1983). Alternatively, the advance may have reached much further until the Altenwald Moraine and the southern part of the Lamstedt Moraine. Which of these scenarios is true, is still unknown (Van Gijssel, 1987; Meyer, 1987). Unlike most other ice-pushed ridges, no glacial basin was found.

6.2.7 Lüneburger Heide

The Garlsdorf and Göhrde end moraines were previously interpreted as Warthian resessional moraines. Meyer (1987) argues that these ridges have already been formed during the Drenthian stage, as they are more or less covered with till. The Barendorf end moraine contains Warthian tills and meltwater deposites and is therefore given a Warthian age (Meyer, 1987). Like the Lamstedt and Altenward ice-pushed ridges, these ice-pushed ridges did not form during the first advance of the ice but they are considered to be recessional moraines formed after the maximal extension of the ice (Meyer, 1983; Van Gijssel, 1987; Van der Wateren, 1995).

6.2.8 Nordrhein-Westfalen

Nijmegen- Krefeld ridges

A discontinuous line of ice-pushed ridges runs between Nijmegen and Krefeld, severely affected by erosion. The ice-pushed ridges on this line can be found in figure 6.11. Between Kleve and Elten the ice-pushed ridge was eroded by the river Rhine (figure 6.26) and were probably formerly connected (Verbraeck, 1975; Van

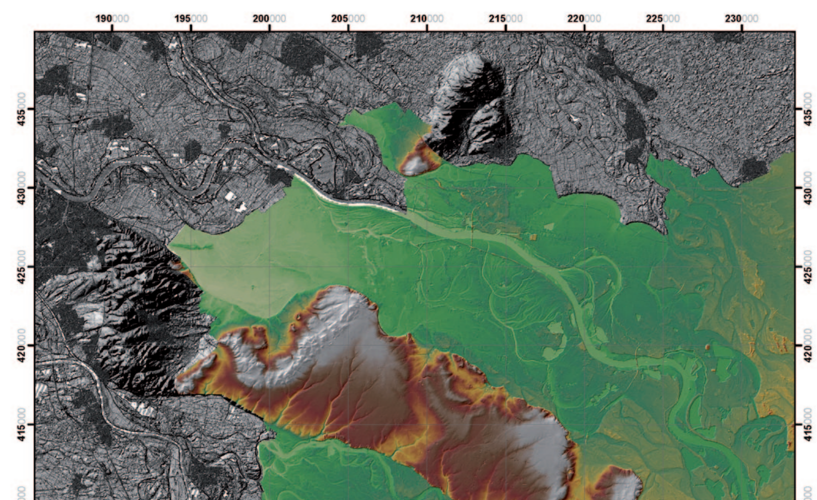


Figure 6.26
Lasar altimetry high resolution elevation image of the ice-pushed ridges near Kleve and Xanten. The ice-pushed ridges and the slowly dipping sandur are clearly visible. From the morphology of some of the ice-pushed ridges the internal structures can be deduced. The ice-pushed ridges and sandurs are sharply eroded by the Rhine after deglaciation. Of the bordering Dutch area only the hillshade image is displayed. The ice-pushed ridge in the westis the Kleve ridge, in the southwestern part the Xanten ridges are presen. The Schaephuyser Hohen are located south of this image and are the southernmost ice-pushed ridges in figure 6.11.

der Meene, 1977). Probably, the Nijmegen ice-pushed ridge was also attached to the Arnhem ice-pushed ridge of the Veluwe. The Montferland ice-pushed ridge is also part of this complex. Strikes indicate a NNW pushing for the northern part. On the eastern side of Montferland pushed glaciofluvial deposits are present (Van der Meene, 1977).

Figure.6.27
Section in the Heyberg ice-pushed ridge near Kleve (from: Klostermann, 1992).

Klostermann (1981; 1997) and Maarleveld (1981) distinguish two phases of ice pushing for the Kleve-Kranenburg ice-pushed ridge based on the morphology of the ridges (chapter 7.5). Besides, two different tills were found separated by meltwater deposits associated with these different phases (Siebertz, 1984). For the ice-pushed ridges Kranenbrug-Xanten, four different ice advance events were separated based on the border of subglacial meltwater deposits and proglacial meltwater deposits. Not every ice advance caused the formation of a ice-pushed ridge (Siebertz, 1984).
South of the Xanten ice-pushed ridges the Bönninghardt ice-pushed ridges and sandur are found. The ice-pushed ridges in this part are very small because they were heavily eroded by the Saalian Rhine (Thome, 1984b). The Schaephuyser Höhen, located just north of Krefeld, is the southernmost ice-pushed ridge in the research area. It consists of Tertiary and Holsteinian sediments. Also Saalian Rhine sediments (Untere Mittelterrasse) and glaciofluvial deposits can be found in the ice-pushed ridge (Thome, 1984a). The internal structure is composed of thrust sheets and folds (figure 6.25). This ridge was most likely formed by one or very few glacier advances from the east (Thome, 1984a). The ridges are severely affected by meltwater erosion, which probably occurred during and after the formation of the ice-pushed ridges. This also caused the formation of the small 'Inselberge' (Zimmermann, 1929), which are isolated small round hills near Krefeld.

6.3 Eskers and tunnel valleys

North Sea
40 km west of Den Helder a large Saalian valley system is present of 40 km long and 7 to 10 km wide (figure 6.28). The beginning and end of this valley were not clearly visible on the seismic records. The valley cuts to ridges interpreted as ice-pushed ridges and incises Holsteinian deposits. The infilling consists of Eemian and Weichselian sediments (Joon et al., 1990; Passchier et al., 2010).

Achterhoek/Twente
In the Achterhoek/ Twente region a system of subglacial channel systems is present (figure 6.28; figure 6.31), which are much smaller than their Elsterian counterparts (Van Rees Vellinga & De Ridder, 1973; Skupin et al., 1993). They were deepened by melt water and probably by land ice. Later they were infilled with glacial outwash deposits and reworked Tertiary deposits (Van Rees Vellinga & De Ridder, 1973). During deglaciation the Rhine could deposit in the depressions. A channel runs from Vreden (Germany) to the SW via Winterswijk to Aalten and Dinxperlo. A second, parallel channel runs from the Nordhorn Basin towards Neede. A third channel cuts the second one near Haaksbergen and runs N-S, aligned with the subcrop of the impermeable Rupel Formation (Dubelaar & Geluk, 1996). It merges with the Vreden-Dinxperlo channel near Winterswijk. The bottom of this system is positioned 70 m below sl. Between Geesteren and Langeveen this system is connected to a several kilometers long gravel ridge, which is considered to be an esker remnant (Maarleveld, 1956; Ruegg, 1983; Van den Berg & Den Otter, 1993). It consists of coarse gravelly deposits with some Scandinavian pebbles. The ridge

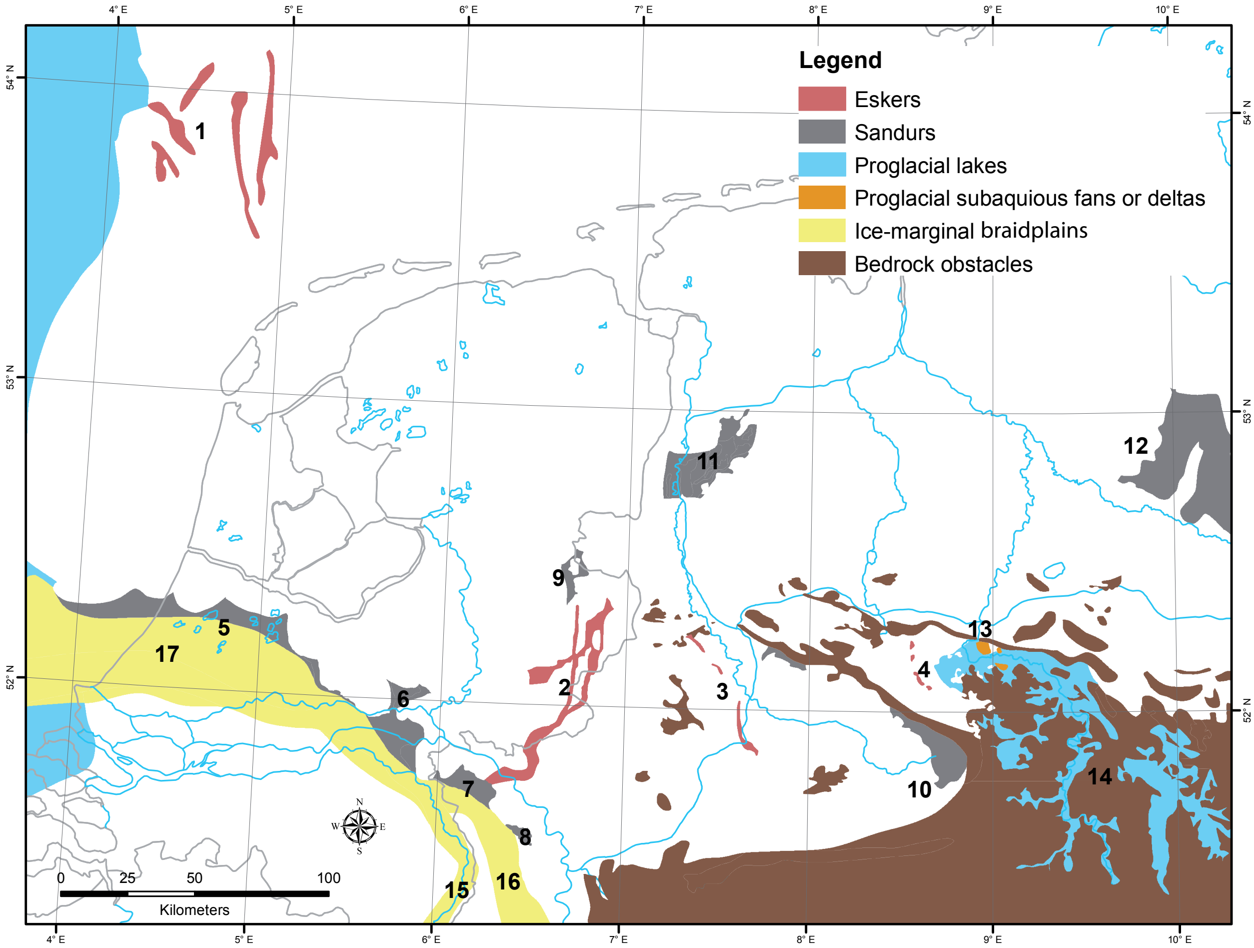


Figure 6.28
Eskers, tunnel valleys and sandurs in the research area; 1= North sea tunnel valleys, 2 = Achterhoek tunnel valleys, 3 = Münster Kieszandzug, 4 = Ravensburger esker, 5 = West/central Netherlands sandurs, 6 = Schaarsbergen sandur, 7= Nijmegen-Kleve sandur, 8 = Bonninghardt, 9 = Dinxperlo sandur (overridden), 10 = Senne 'sandur', 11 = Hümmling (overridden), 12 = Lüneburger Heide (formed during readvance), 13 = Porta Westfalica subaqious fans, 14 = Lake Weser, 15 = Meuse, 16 = Rhine, 17 = Rhine-Meuse ('Unit S4').

is rather wide compared to other known eskers (Rappol, 1993a).
Dubelaar & Geluk (1996) suggested that these two channel systems may be linked to two different glaciation phases with different flow directions.

Münsterland Embayment esker
In the Münsterland Embayment a large elongated filled up tunnel valley is present called the Münster Kieszandzug (figure 6.28 - Thome 1983/4?; Skupin et al., 1993). The feature incises several tens of meters into the Cretaceous limestones (figure 6.9 Driesen, 1990), it is approx. 80 kilometres long and one kilometre wide (Klostermann, 1995). The bottom part is filled in with coarse sand and gravel, the top with sand. The northern part has a NNW-SSE orientation and the southern part a more NW-SE orientation, which roughly reflects the flow direction of the ice. According to Skupin et al. (1993) this feature was formed between an active and a dead ice body.

Ravenburger Kiessandzug
Between the Wiehengebirge and the Teutoburgerwald a 22 km long ridge is present, which is called the Ravensberger Kiessandzug. The ridge is orientated N-S and NNW-SSE (figure 6.28). (Skupin et al., 2003) attributed the shape and orientation of the ridge to the collision of two ice streams, the Aue-Hunte glacier and the Porta-glacier. Thome (1983) interpreted the ridge as a subglacial meltwater ridge, like the one in the Münsterland Embayment.

6.4 Sandurs

6.4.1 North Sea

In the North Sea glaciofluvial deposits from sandurs have hardly been found. It is well possible they have hardly been formed in this region, given that the ice front may have bounded the proglacial lake North Sea. Under these conditions subaqueous till and outwahs deposits would form rather than sandurs or kames. Deposits that were formed were amongst others, eroded during the Eemian transgression (Joon et al., 1990; Laban, 1995). These would have occupied low spots, and have been subject to post-Drenthe dissective erosion.

6.4.2 The Netherlands

In the Netherlands some sandurs are present that were formed as large outwash fans in front of the ice margins, usually marked by ice-pushed ridges. Good examples are the sandurs on the western side of the Utrecht Ridge, in Twente and in the Southern part of the Veluwe (figure 6.28). In many cases, pushed outwash deposits are present in the most proximal part of the sandur (Ruegg, 1983; Bakker, 2006).

The sedimentology of the sandur deposits in the Netherlands was extensively described by Ruegg (1977; 1983). In the Netherlands these deposits are called the Schaarsbergen Member of Drente Formation (Westerhoff et al., 2003; Bakker et al., 2003).

The Central and Western Netherlands

As mentioned before, the western edge of the Utrecht Ridge is marked by large sandurs. These meltwater deposits were formed by large meltwater streams from the melting ice lobe. The meltwater eroded into the ice-pushed ridges forming large meltwater channels. The sandurs formed in front of these channels, sometimes in the shape of an alluvial fans (Ruegg, 1983). The large sandur of Soesterberg was described by Augustinus & Riezebos (1971), here the sandurs has a slope of approximately 2m/km. In the Gooi region also sandur and kame deposits occur (Ruegg, 1977; 1983; Ruegg & Koopman, 2010). Busschers et al. (2008) presents OSL-dates from the superelevated Rhine braidplain 'S4' that connects to the Utrecht Ridge sandurs, at Utrecht De Uithof and downstream, yielding ages of 130-157 ka. Yet unpublished OSL results obtained from high-elevated glacio-fluvial deposits with modest ice-thrust deformation (Schaarsbergen; Arnhem-Burger's Zoo - Pawley & Busschers, pers. comm.) are in agreement with the OSL ages obtained for the Utrecht ridge (155-145 ka).

The Eastern Netherlands

In Twente, west of the Ootmarsum ice-pushed ridges, the Bruinehaar-Dinxperlo sandur can be found. These were formed when the ice margin was located at the Rehburg line, probably from the ice lobe in the Nordhorn Basin (Van den Berg & Beets, 1987). Unlike the outwash fans in the central Netherlands this outwash fan is covered by a basal till formed during the overriding by the ice (Van den Berg & Den Otter, 1993).

6.4.3 Germany

In Germany, meltwater deposits are very abundant, the ratio between tills and meltwater deposits is 1:3 (Ehlers & Grube, 1983). In northern Germany these meltwater deposits are considered to be remnants of large outwash plains, which can be divided into deposits forming in front of the ice (Vorschüttsande) and during ice retreat (Nachschüttsande). The Vorschüttsande are deposited in thick outwash plains and usually covered with tills (Caspers et al., 1995), whereas the Nachschüttsande are usually only formed in (deglaciation)channels. According to Ehlers & Grube (1983) this can be explained by the fact that dead ice was unable to produce larger amounts of meltwater that an active ice front. This is unlikely, because a dead ice body would produce more meltwater than prograding ice because dead ice is much weaker and it will thus melt more easily without being replaced by new ice. Probably, the deglaciation channels were present at the same locality for a longer time whereas the ice marginal rivers. Most likely, these Vorschüttsande do not only represent sandur deposits, but also deposits from ice marginal rivers (chapter 6.5).

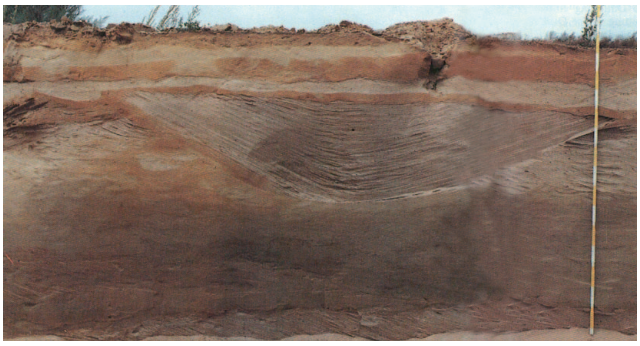


Figure 6.29
Erosional channel at the top Bönninghardt sandur (from: Klostermann, 1988).

Hümmling
The Hümmling region is located on the western edge of the Cloppenburg Geeste. In the Hümmling region both SSW-NNO and NO-SW trending heights and depressions occur (chapter 6.1 - figure 6.6). These two directions can also be seen in the neighbouring areas, for example in Ostfriesland (ch.. till Lower Saxony.). The highest point is the Windberg (73m asl). The relief can be linked to ice flow direction (Schröder, 1978). They may have formed partially directly by the ice but also by meltwater.
The Hümmling mainly consist of thick meltwater deposits from a sandur that formed during the advance of the ice front. The base of these deposits are glaciolacustrine deposits. They consist of fine sands due to the lack of clay in the source area (Schwan & Kasse, 1997). They argued that a proglacial lake formed in local depressions inherited from the Elsterian. The lake was fed by glacial meltwater or deflected ice-marginal rivers. In the meltwater deposits a range of flow directions was found, that may reflect irregularities in the ice lobes of the irregular ice margin (Schwan & Kasse, 1997). On top of this a small till layer is positioned and a deflated gravel horizon, comparable to 'keizand' in the Netherlands. Together with the striking relief this implies that the sandur was overridden after its formation by SW to SSW flowing ice (Schröder, 1978). This was confirmed by fabric measurements (see chapter 6.1). Fabric and glaciotectonic evidence showed that the Hümmling sandur was also somewhat ice-pushed from the NE (Schröder, 1978). Most caused by the same ice stream as the ice stream that shaped the Hondsrug (Rappol, 1991a).
On the age of the glaciofluvial deposits in the Hümmling some discussion exists. Zandstra & Skupin (2006) found a large amount of erratics from the area around Oslo (area IV, 10), which made them conclude that the meltwater deposits are of Elsterian age. Schröder (1978) states that only the lower section of the meltwater deposits has an Elsterian age, Schwan & Kasse (1997) conclude that all deposits in the Hümmling have a Saalian age.

Lüneburger Heide
In the eastern part of the research area the large sandur of the Lüneburger Heide is present. It is thought to have formed when the ice retreated far to the northeast as it is positioned above till deposits (Ehlers et al 2004; LBEG, 2006).

Senne Sandur
In the southeastern part of the Münsterland Embayment a large glaciofluvial features has been interpreted as a large sandur (Hoyer & Vogler, 1977; Klostermann, 1995). These sediments overlay tills, and were previously interpreted as a kame by Seraphim (1979) and as a sandur by Klostermann (1995). Most likely, these deposits are glacio-deltaic deposits formed in the Münsterland Embayment deglaciation lake (Winsemann & Meinsen pers comm.).

Nijmegen-Krefeld
Two large sandurs are present in this area Reichswald and Bönninghardt (figure 6.AD). At some

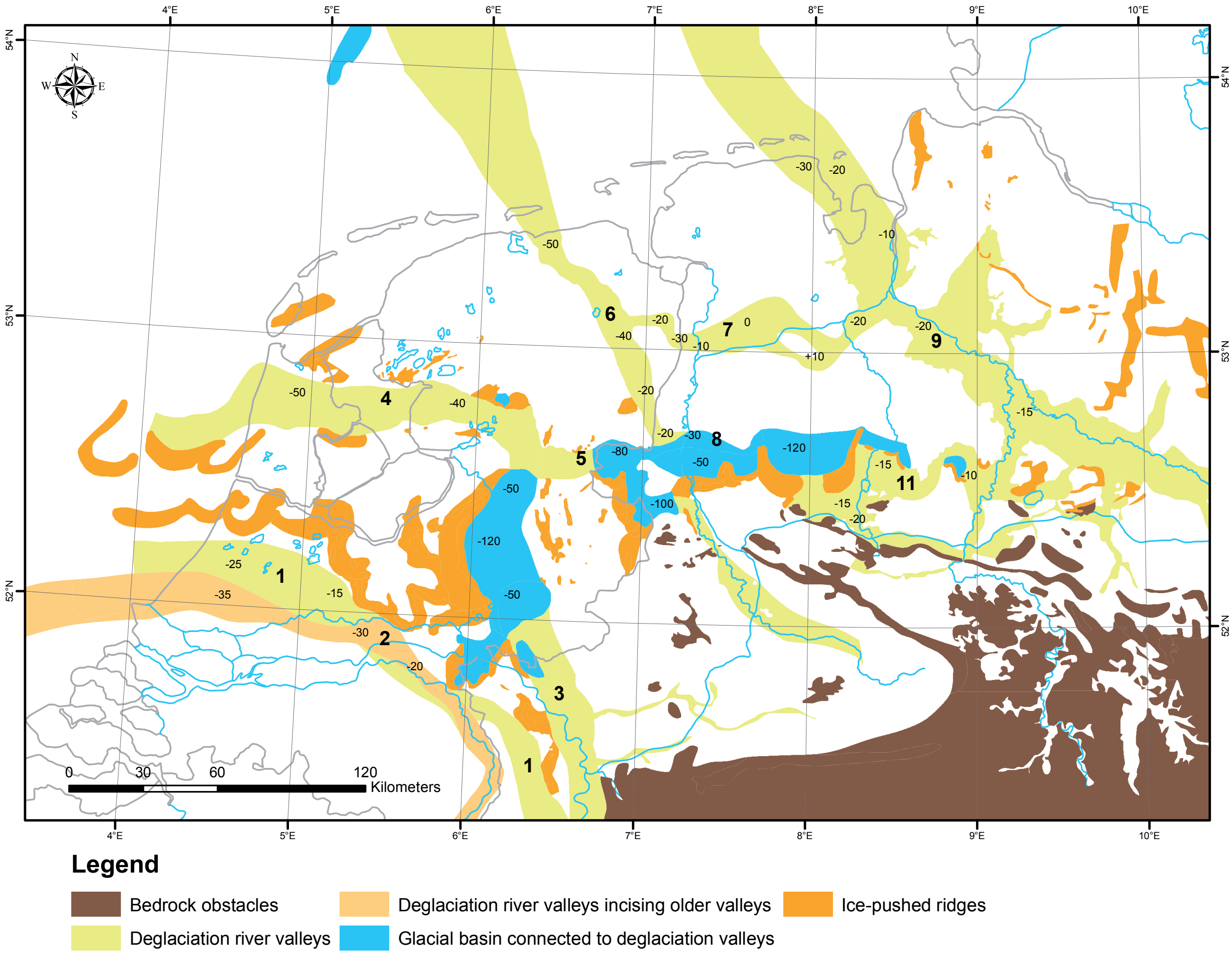


Figure 6.30
Deglaciation river valleys (deglaciation phase) and ice-marginal river valleys (maximum extension). The Rhine terrace is not considered to be a deglaciation channel, but an ice-marginal river in which a deglaciation river incised. The base level of these valleys are indicated, after: Ter Wee (1983); LBEG (2006); Busschers et al. (2008). 1 = The ice marginal Rhine and Meuse ('Unit S4'), 2 = Deglaciation incising Meuse ('Unit S5'), 3 = Deglaciation Rhine, 4, 5 = Vecht Valley, 6 = Hunze Valley, 7 = Easter Hunze Valley (Leda-Jümme depression), 8 = Hase depression, 9 = Weser-Aller valley, 10 = Elbe valley, for the glacial basins see figure 6.32.

locations these sandurs are still attached to ice-pushed ridges, however, most of these moraines have been eroded by the Rhine after the Saalian (Klostermann, 1997). The Bönninghardt complex consists mainly of a sandur plain (figure 6.29). In this sandur a lot of glauconite can be found that originates from the reworked Tertiary deposits in the adjacent ice-pushed ridges and adjacent hinterland of the Münsterland Embayment (Thome, 1984b; Klostermann, 1988).

6.5 Ice-marginal rivers and deglaciation rivers

During the onset of the MIS6 glaciation the course of large rivers such as Rhine and Weser were deflected parallel to the ice margin flowing to the west. During the deglaciation large amounts of meltwater was drained from the glaciated area, in some cases causing the formation of deeply incised valleys: Hunze, Vecht and Elbe Valley (figure 6.30). At several locations the deglaciation rivers entered large depressions filled with melt water (in most cases a glacial basin). These lakes were nodes in the valley network. They trapped sediments, but allowed the water to spill onwards ('cascade' - chapter 5.6).

6.5.1 Pradolinas during the onset and maximal extension of the glaciation

Ice-marginal rivers and sandur will have existed also in lines in front of the prograding ice front, before the maximal ice extent was reached. The 'Vorschüttsande' have not only been built up by sandurs as advocated in the German literature (chapter 6.4), but also by ice marginal river deposits. In the Netherlands these kind of deposits have often received special attention. In Drenthe, the deposits under the Saalian till are part of the Peelo Formation, which is generally regarded to be as a whole of Elsterian age (Bosch, 1990; Westerhoff et al., 2003). However, the top of the Peelo Formation contains subunits of coarser sands which could very well be attributed to Saalian ice-marginal rivers reworking Elsterian sediments.

Rhine and Meuse

The youngest unit of the Rhine deposition in the central Netherlands is extremely coarse and very wide spread (figure 7.11). Before it was incorporated in the ice-pushed ridges, it was a wide fan merging with ice-marginal rivers from the east ('Unit S3' - Busschers et al. (2008) - chapter 7.4.2). An OSL-date was performed in the Utrecht Ridge yielding a an age of 168±19 ka, MIS6. In these deposits downstream fining can be observed, the base level of this system is estimated at -30m and -10m when the north Sea lake was a at high stage (Cohen, pers. comm.). During the maximal extension the Rhine and Meuse acted as an ice marginal river depositing 'Unit S4' (Busschers et al., 2008). Units S3 and S4 in series indicate that the Rhine was gradually forced into the final position during the maximal ice extension.

South of the ice front in the Northern Netherlands (phase 1; Rappol; phase 2; Busschers) most likely an ice marginal river was present in the current Vecht valley (Busschers et al., 2008). This river drained meltwater from the east, including the Weser catchment (Klostermann, 1992), finally it reached the Rhine. This happened when the ice front was present along the Northern Netherlands and the Rehburg line, forcing all the river and meltwater to flow towards the west (Figure 7.11).

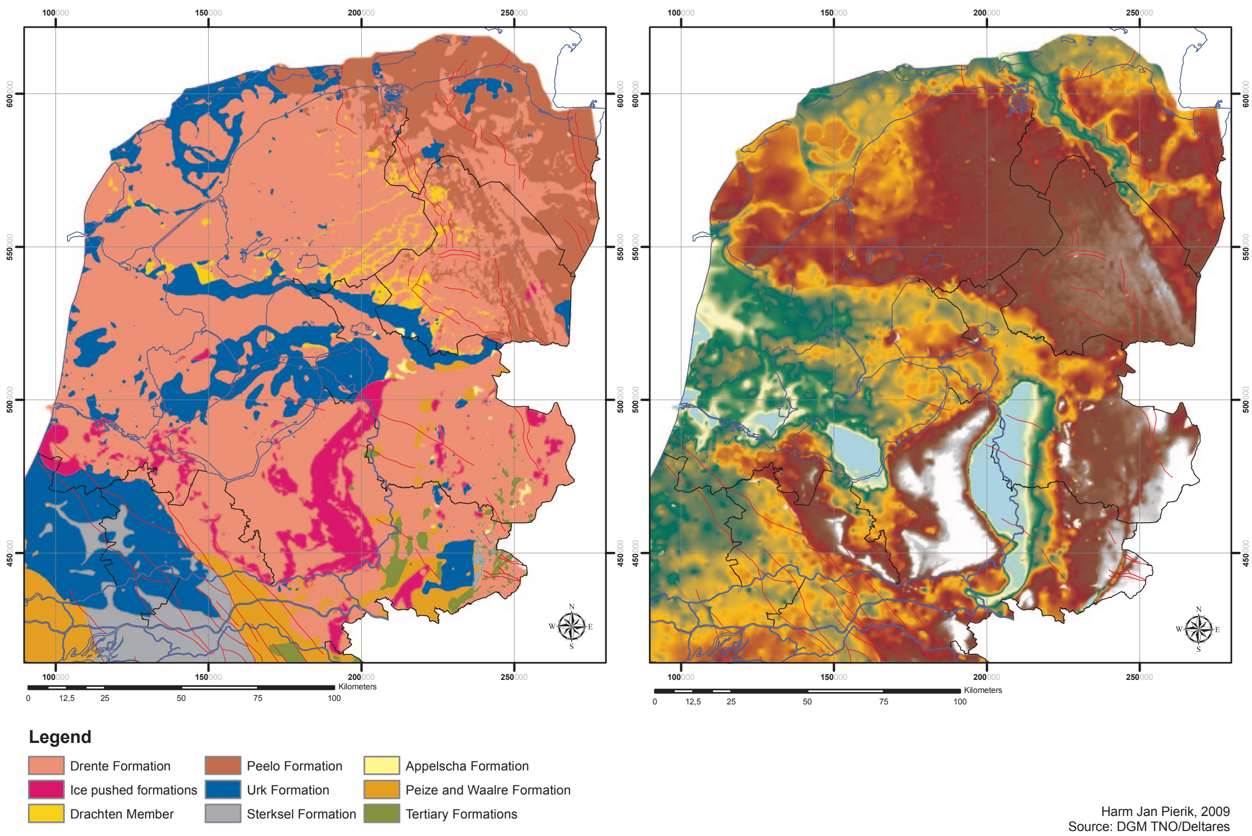


Figure 6.31
Extension of the Early and Middle Pleistocene deposits, including the Saalian ice-pushed ridges and Drenthe Formation, consisting mainly of tills and meltwater deposits. The current elevation of the top of these deposits can be seen in the right image. The glacial basins and deglaciation valleys of the Vecht and Hunze are clearly present, in the Achterhoek the tunnel valley system can be seen. In the Northern Netherlands Holocene marine erosion took place into these deposits (after: DGM, 2009).

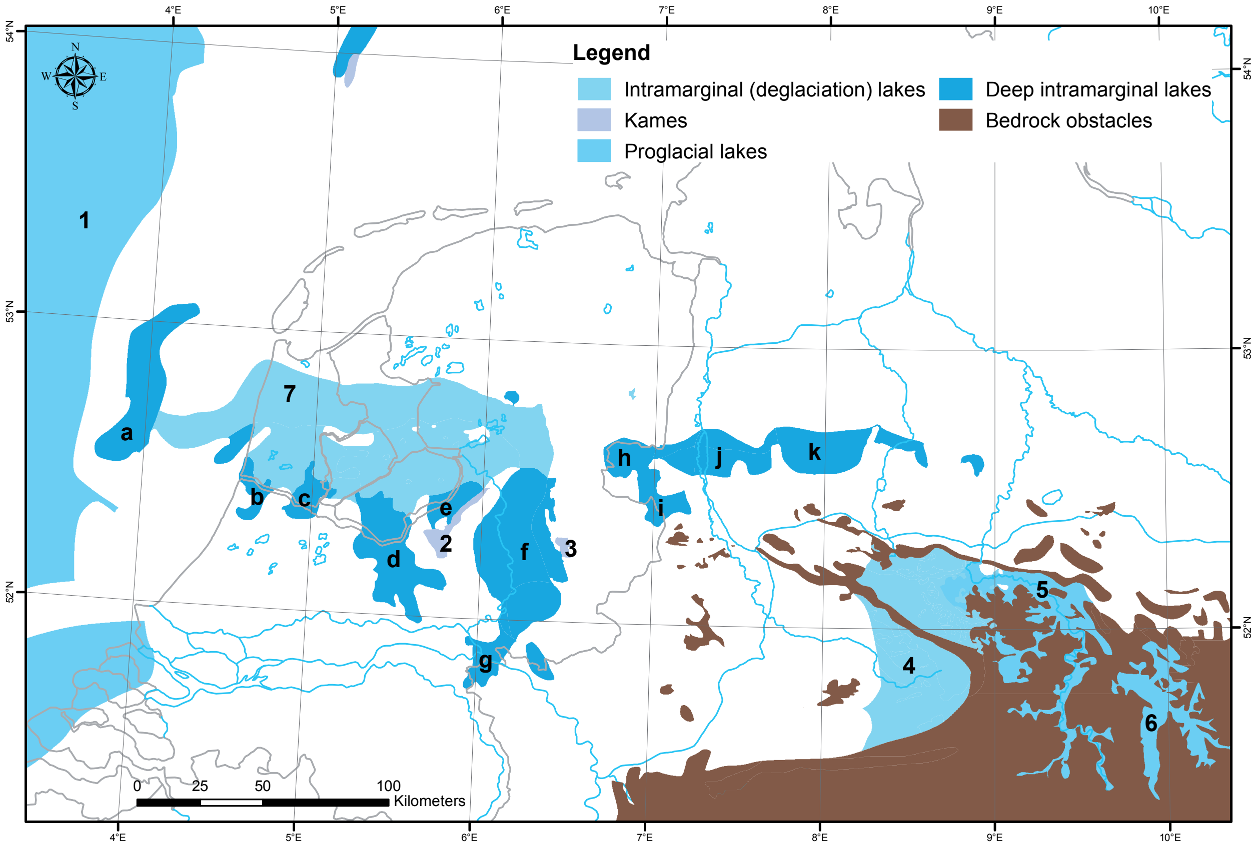


Figure 6.32
Extra-marginal glaciolacustrine sediments and morphological features. Extramarginal lakes 1 = North Sea proglacial lake, 2 = Kame Leuvenmesche Beek, 3 = Markelo kame, 4 = (deglacial) Lake Münsterland, 5= Lake Weser, 6 = Lake Leine. Intramarginal (deglaciation lakes): 7 = Holland Lake, a = North sea, b = Haarlem Basin, c = Amsterdam Basin, d = Geldersche Vallei, e = Nunspeet Basin, f = IJssel Basin, g = Valburg/Kranenburg basin h= Wilsum Basin, i = Nordhorn Basin, k = Quackenbrück Basin.

6.5.2 Deglaciation rivers

Rhine and Meuse

Right after the deposition of 'Unit S4', a deep valley was formed, that reworked the southern edge of these deposits. The deposits marking this valley are known as 'Unit S5', and have their base at -35m msl. No accurate dates could be obtained from these deposits. Its architectural position indicates a drop in the lake level of the proglacial lake North Sea of approx. 20 m. in the beginning of deglaciation. This 'S5' valley hosted the Meuse during the rest of the deglaciation. The Rhine entered the deep IJssel basin (Van der Meene, 1977; Busschers et al., 2008 - chapter 6.7.3), which overspilled towards the continuation of the Vecht deglaciation valley.

Vecht and Hunze

It was suggested by Ter Wee (1962; 1966; 1979) and Jelgersma & Breeuwer (1975) that the river Vecht and Hunze formed as ice-marginal rivers along the ice margins in the northern Netherlands. This happened after the maximal ice extension occurred in resp. phase E and F (chapter 7.2). However, no tills were found in these valleys and no ice contact fans are attached. Besides, the valleys are deeply incised in the surrounding Saalian sediments. This is considered uncommon for ice-marginal rivers in the direct vicinity of the ice front, as they typically deliver a lot of sediment, which reduces the ability to erode. Excess incision depth could indicate loss of bedload supply downstream in through-passed lakes in the glacial basins (Van den Berg & Beets, 1987).

Bosch (1990) attributes the origin of the Hunze valley to a narrow glacier tongue flowing to the SSE,

after which meltwater erosion occurred. The Hunze valley may have formed after overflow of a glacial basin from the Rehburg line, in which the river Ems drained (Van den Berg & Beets, 1987). In this view the Vecht and Hunze are deglaciation valleys rather than ice-marginal river valleys.
On the figure 6.31 also an eastern part of the Hunze valley can be seen with a deep bottom level. This may indicate that this part of the Hunze valley was longer active during the deglaciation, but it could also be the consequence of marine erosion during the Eemian. In Germany this channel can be traced between the Oldenburg and Ostfriesland till plateaus called the Leda-Jümme depression by (Speetzen & Zandstra, 2009). This 60 km long and 20 km broad valley was not as deeply incised has its base appox. around current sea level. In (LBEG, 2006) was deduced that Weichselian fluvioglacial deposits are mostly directly positioned on Elsterian deposits. This indicates that no deep incision took place in this valley during the Saalian deglaciation.

Like the Hunze, the Vecht valley system incised deeply, probably due to the same reason (Van den Berg & Beets, 1987). The IJssel basin was the most important lake from which the Vecht was drained, other basins are the Wilsum and Nordhorn glacial basin lakes. South of Texel-Wieringen coarse sands and gravels from this deglaciation river are present between -60 and -40 NAP in a broad meltwater valley system (De Gans, 1994).

Weser-Aller
The Weser-Aller valley can be traced in the present landscape as a large elongated depression of several tens of kilometres wide (figure 6.H). It was formed during the deglaciation of the Drenthe stage when it drained a substantial part of the eastern part of the research area. It was enlarged when it became a ice marginal river during the Warthe readvance stage (Meyer, 1983; Ehlers et al., 2004). This river drained the whole ice front up to Poland into the north sea (Fig 3.13b).

Other deglaciation rivers
North of the Wiehengebirge a complex of meltwater stream deposits is present. Meltwater that was left in the Weserbergland crossed the Porta Westfalica and headed north in the Aller-Weser valley. Some of the valleys were formed between some Cretaceaous outcrops and ice-pushed ridges as major obstacles. Eventually, this water flew to the Quackenbrück basin were it finally drained trough the Vecht and Hunze. In the Münsterland Embayment a major Weichslien valley is present along the Teutoburgerwald, this system was probably already active during the Saalian when a deglaciation river ran to the northwest and eventually into the Nordhorn Basin.

6.6 Extra-marginal lakes and kames

Large kames or kame terraces are rare in the research area; and especially in NW Germany (Ehlers & Grube, 1983). Yet, large proglacial lake areas are suggested to have covered the area during the maximum extent of the Saalian. They were formed at the ice-sheet front, and were fed by the meltwater outflow and by local rivers draining the southern mountain range (figure 6.32).

6.6.1 Glacial lake Weser
In the southeastern part of the study area large proglacial lakes have existed (Klostermann, 1992; Van der Wateren, 1994; Winsemann et al., 2003; 2004; 2009; 2010). The glacial lake Weser reached a stage of 200 m above sea level and it extended up to 50 km southward to Göttingen. It spilled into the Rinteln Lake (Winsemann et al., 2007).
South of the Wiehengebirge several large subaqueous fans and deltas are recognized which were previously mapped as fluvial deposits and kames (Winsemann et al., 2007). The best investigated one is the Porta Sedimentary Complex (Hornung et al., 2007; Winsemann et al., 2004; 2009), southeast

of the Porta Westfalica bedrock gap. This fan complex is about 10 km wide, and it extends over 15 km to the SE. The total thickness of the Porta Complex is 40–72 m. It overlies glaciolacustrine deposits and tills. The sedimentological characteristics of the Porta Westfalcia pass delta has been described extensively by Hornung et al. (2007) and extended by Winsemann et al. (2009) with a 3-D model of the fan. Two other fans are the Emme Delta and the Coppenbrügge subaqueous fan. The Emme delta is thought to reflect a relatively stable ice margin position whereas the Coppenbrügge fan reflects a retreating ice margin (Winsemann et al., 2004). In the latter, evidence of glacier advance over the fan has been found (Winsemann et al., 2004).

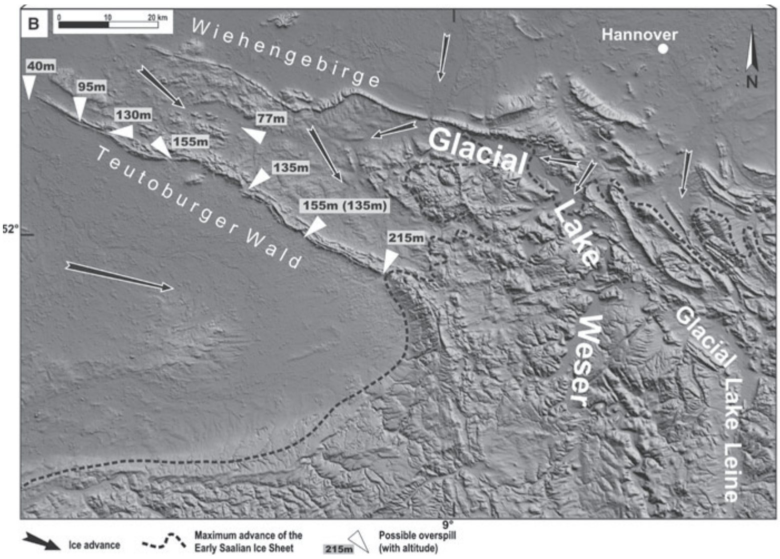


Figure 6.33 Locality and basal height of the overspills in the Teutoburgerwald (from: Winsemann et al., 2009).

From the architecture of the fans the lake level history could be reconstructed. The level of this glacial lake was initially at an altitude of some 55 m asl but subsequently rose to a highstand at approx. 175 m asl. Ice rafted debris in the upper Weser Valley indicated an even higher lake level of 200 m asl. After this rise two stages of lake level drop are interpreted (Winsemann et al., 2009). The first lake level drop amounted 40 to 60 m, and is thought to have been generated by overspilling over the Teutoburgerwald (base: 135m) (figure 6.33). Subsequently, Gilbert-type deltas formed on the truncated fan margin, recording a second lake-level drop in the range of 30 to 40 m. The water could flow away through the more westerly located overspills with a base of app. 90m. The lake level drops in the Lake Weser and the spills in the Teutoburgerwald may be associated with ice front retreat and the formation of Lake Münsterland Embayment (Winsemann, Meinsen pers. comm.).

6.6.2 Proglacial lake North Sea

In large parts of the Dutch North Sea varved glaciolacustrine deposits are found, known as the Cleaver Bank Formation (Cameron et al., 1986; Joon et al., 1990: Cameron et al., 1992; Laban, 1995) Their wide occurrence and considerable thickness (4-6 meters, 29 in the Northern Dutch part) indicates either a high amount of sediment supply or a long ice front stagnation at its maximum (Laban, 1995). As these deposits are not covered by till and occur below Eemian marine sediments (Laban, 1995; Busschers et al., 2008), deposition during the glaciation is plausible. During the Eemian transgression these deposits were severely affected by erosion (Laban, 1995). Major erosion also occurred during the Weichselian, when large river valleys were present in this area.

It has been suggested by several authors that in the southern North Sea area a large proglacial lake formed during either the Elsterian or the Saalian glaciation, or at both times (figure 6.32 – Belt, 1874; Smith, 1985; Gibbard, 1995; Gibbard, 1988; Meijer & Preece, 1995; Gupta et al., 2007; Busschers et al., 2008). The (melt) water transported by the ice-marginal and periglacial rivers (Rhine, Meuse Thames and Elbe), was stored in this lake. The southern edge was hold up by a bedrock barrier consisting of chalk (Weald Artois) across the (present) Dover Strait and the adjacent Tertiary deposits to the north of it. In the north, the lake was dammed of by the Fennoscandian and the British ice sheets, implying that those two ice sheets must have been in contact. Busschers et al. (2008) inferred a water level of the lake at the time of Saalian glaciation maximal extent that was close to interglacial high-stand sea

levels. They based this on the position of the ice marginal Rhine deposits ('Unit S4') that built out into in this lake, supported by the depth of the Cleaver Bank Formation, and applying corrections for post-Saalian subsidence.
According to Toucanne et al., (2009) first spillage of water over and the start of erosion of the Dover Strait barrier occurred at 170 ka. Around 155 ka, probably coinciding with the major deglaciation, this discharge significantly increased (Toucanne et al., 2009). After deglaciation, the water drained to the north. This implies that large parts of the North Sea must have been free of ice at that time, which most likely corresponds to the Warthe substage (figure 3.11b - Toucanne et al., 2009). First tests of numerical modelling done by Van Hoesel (2009) support the idea that erosion due to spillage from the lake was non-catastrophic and must have taken 8.5 to 15 ka, centred around the maximum ice extent, to explain invoked amounts of topographic lowering of the Dover Strait.

6.6.3 Kame terraces of the Veluwe and the Eastern Netherlands

On the northern edge of the Veluwe several kames were found (Maarleveld, 1953; 1981). The largest one is the Leuvenumsche beek valley kame. It was considered as deposits formed in an ice marginal lake (kames) because of the rhythmic variation in grain size (not necessarily varves) and relatively high percentages of Scandinavian gravel (Maarleveld, 1955; Ruegg, 1977). The finest deposits occur in the middle part of the valley, in front of the presumed ice front (Postma, 1997). They consist of very heavy clay (Edelman & Maarleveld, 1958). At the flanks deposits of subaquous sediment gravity flows were found, which originated from the adjacent ice-pushed ridges (Postma et al., 1983; Postma, 1997). The ice lake was dammed off by the Veluwe ice-pushed ridges on the west, east and south and the ice on the north. The bottom of the infilling lake deposits is positioned at -50 m NAP and shows a U-formed shape. Despite this, the ice probably never intruded the region of the Leuvenemsche Beek kame (Postma, 1997). The kame formed simultaneously with the ice pushed complex. Another kame formed between ice-pushed ridges is present near Markelo (Rappol, 1993a).

6.7 Glacial basins and intramarginal lake deposits

Glacial basins are morphological features that form after the formation of ice-pushed ridges. In these glacial basins deep intramarginal lakes form during glaciation. The origin of the basins is not only related to glacial surge as suggested by Jelgersma & Breeuwer, (1975), but is probably also related to the differential hydrostatic pressure and the occurrence of subglacial channels (De Gans et al., 1987). Probably, the basins initiated by subglacial meltwater processes and were later on overdeepened by glacial scouring processes (De Gans, 1994). In the Netherlands, large glacial basins are associated with the southern fringe of the ice margin at maximal extent. Their dimensions are summarized in table 6.2. From the table can be seen that the size of the glacial basins increases towards the east. The volume of the adjacent ice-pushed ridges increases proportionally (De Gans et al., 1987). The basins are formed in the part of the North Sea basin that contained rather coarse middle Pleistocene fluvial deposits (chapter 3 - Van den Berg & Beets, 1987).

Glacial basin	Length (km)	Width (km)	Depth (m)
Beverwijk	?	?	-112?
Haarlem	15	10	-120
Amsterdam	25	15	-125
Gelderse Vallei	50	20	-130
IJssel valley	90	25	-140

Table 6.2 Glacial basins and their dimensions (from: De Gans et al., 1987).

The glacial basins were filled up during the deglaciation. Most basins were infilled with sediments from multiple sources. A major source for sediments were the unstable slopes of the adjacent ice-pushed ridges. In some basins a major river deposited coarse bedload in a fan. This happened in the IJssel basin that was mainly filled up with sediments delivered by the Rhine. In the distal parts clayey deposits are found, called the Twello Member of the Kreftenheye Formation (Westerhoff et al. 2003; Busschers & Weerts, 2003). These deposits can also be traced in the basins towards the west. Here also, finer materials were deposited from local rivers (De Gans et al., 1987; 2000). These deposits are called the Uitdam Member of the Drente Formation (Bakker et al., 2003).

6.7.1 Basins in the Dutch North Sea sector

Laban (1995) found two glacial basins (figure 6.9). A 20 km long SW-NE trending valley 50 km north of Terschelling was found. In the southern part of the valley lodgement till was found that serves as evidence for a glacial origin. The ice probably filled an Elsterian tunnel valley. A second basin was found west of the coast of Noord-Holland. Unfortunately, no erratics were analysed and no fabric measurements were done, therefore the push directions of the ice are difficult to reconstruct for the North Sea. These basins were filled with varved lacustrine clays (Cleaver Bank Formation).

6.7.2 Noord-Holland basins and the Holland Lake

In the province Noord-Holland several glacial basins are present. The main basins are the basins of Haarlem and Amsterdam (figure 6.AK). Under the Vecht deglaciation valley the small Spanbroek glacial basin occurs. The NE-SW orientation of these basins indicate that they were formed by ice flowing from a NE direction (De Gans, 1994).

The Amsterdam basin is the best described glacial basin in the research area (e.g De Gans et al., 1987; 2000). A cross-section trough this basin can be seen in figure 6.34. Because the other basins in this region appear to be quite similar, the Amsterdam basin serves as a good example. At the bottom of the basins a discontinuous layer of till was found suggesting severe meltwater erosion during its formation. This is confirmed by the fact that thick layers of glaciofluvial sands were occasionally found beneath the tills (Van den Berg & Beets, 1987). The major part of the infilling of these basins consists of glaciolacustrine clays and silts and glaciofluvial sands (De Gans et al., 1987; 2000; Beets & Beets, 2003). The lacustrine sediments were probably provided by sediments from the Rhine and small rivers that drained in the lakes during the deglaciation. They probably eroded the northern edges of the basin (figure 6.31) and deposited clays derived from the tills just north of the glacial basin (De Gans, 1994). These fine grained deposits interfinger with coaser grained deposits formed by mass wasting of the ice-pushed ridges. After the Eemian transgression marine clays were deposited in the depressions (figure 6.34).

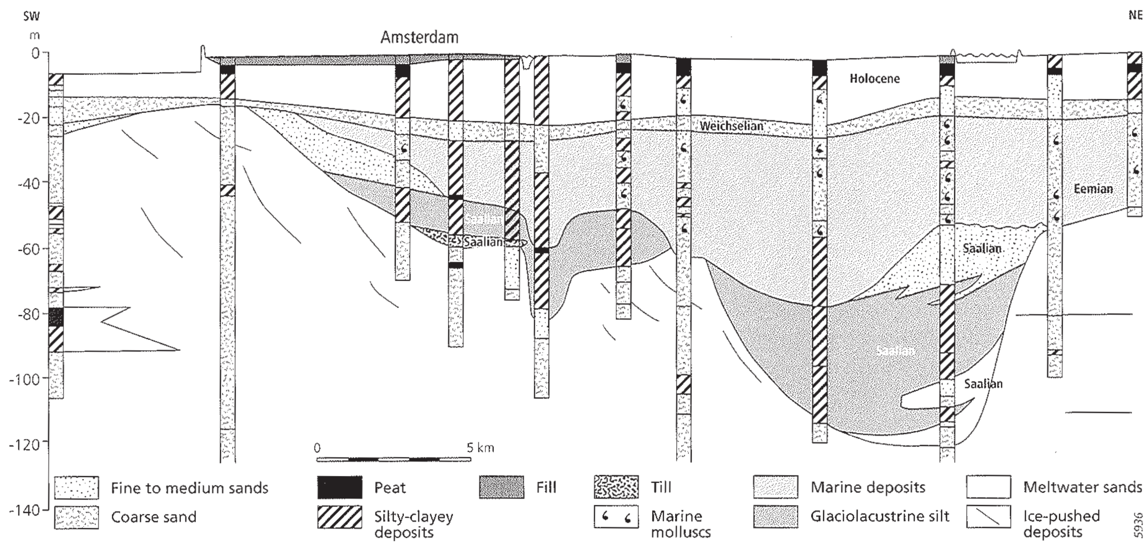


Figure 6.34
Cross-section through the Amsterdam glacial basin, the vertical scale is x100 exaggerated. From Ehlers (2005 - Fig 3.11) after De Gans et al. (1987).

Holland Lake
Beets & Beets (2003) suggested that the subglacial basins and the area between those basins formed a large interconnected lake (Holland Glacial Lake) during the deglaciation. It covered an area of at least 3500 km^2. The edges were formed by the ice-pushed ridges in the south and a not yet isostatically adjusted peripheral bulge in the north. Its western extension was situated in the North Sea, probably close to the western limit of the Saalian ice cap. De Gans et al. (2000) suggested that the lake was separated from the North Sea here, by a sill at a depth corresponding to the estimated lake level (25–30m below NAP).
According to Busschers et al. (2008) the Holland Lake could very well reflect the (terminal) expansion of a much larger proglacial lake that covered a large part of the North Sea (see chapter 6.6).
Approximately 1 ka after its formation the Holland lake drained and the lake level was significantly reduced. This left shallow pools in the deeper, central part of the not yet filled depression. Beets & Beets (2003) estimated that the time interval between the earliest lake sediments in the Amsterdam Basin and the Eemian highstand took about 5–6 ka. Especially the northern half of the Holland Lake was eroded by incision of the Rhine and erosion by wave- and tide-induced currents during the Eemian transgression (Beets & Beets, 2003).
The deep lake and pool deposits were left undisturbed. In the early Eemian, the pools in the centre of the depressions developed into larger lakes due to the higher groundwater level caused by the rising Eemian Sea (Beets & Beets, 2003).

6.7.3 The IJssel Basin
The IJssel glacial basin was formed at the location of the Holsteinian and Early Saalian Rhine and the Gelderse Vallei is located on the early Saalian Meuse valley (Van den Berg & Beets, 1987; Busschers et al., 2008 – see chapter 3.5). The IJssel valley is strongly asymmetric with a steep flank on the ice-pushed ridge side (near Apeldoorn) and a gentle slope towards the east. This flank coincides with the border of the Zuiderzee basin and is the dipslope of the late Tertiary marine sediments. The IJssel Basin probably continued in the southeast into Germany (figure 6 AK). Here, fine sandy basin infilling ('Beckenschluff'- Twello Member equivalent) was found by Klostermann (1992; 1995). These fine deposits originate from the Rhine and were formed in quiet water.
The IJssel basin was mainly filled in with fluviolacustrine sediments and sediments from the Rhine (De Gans et al., 2000; Busschers et al., 2008). The Rhine dropped its bedload in the lake, most of the suspended load passed a sill in the north of the IJssel Valley Lake at a depth of about 25m below present NAP (NAP = Dutch Ordnance Level sea level). Westward beyond this sill, the water and suspended sediment entered the Holland Lake (Beets & Beets, 2003). After the major drop of the water level in the Holland Lake and isolation of the glacial basins, the water was drained through the Vecht valley were it could incise deeply due to the lack of sediment.

6.7.4 Glacial basins of the Rehburg line
In Germany, typical glacial basins (German: Zungenbecken) are associated with the ice-pushed ridges of the Rehburg Line. The deepest one is the Quakenbrück basin (120 m - Meyer, 1987). It is located north of the Fuerstenauer Berge of the Rehburg line. The basin has an asymmetrical shape (LBEG, 2006) with -130 m Vechta, 100-200 m south of the Dammer Berge 100-150 Emsland area (Van der Wateren, 1995). The infilling consists of glaciofluvial sands and finer grained 'basin deposits', see figure 6.35. In the profile it can be seen that the top of these Saalian deposits is located approx. 50 m below the current surface. This means that this basin was most likely still a lake at the end of the deglaciation, i.e. it was not completely filled up by sediments from the deglaciation rivers.
On the eastern side of the Uelsen-Oldenzaal ridge the Nordhorn basin occurs (Kluiving, 1994). The deepest part of the basin is located closest to the ice-pushed ridge, 90 m near Uelsen and 130 m near Nordhorn. The southern edge of the basin, near Nordhorn, is a very steep contact with Cretaceous

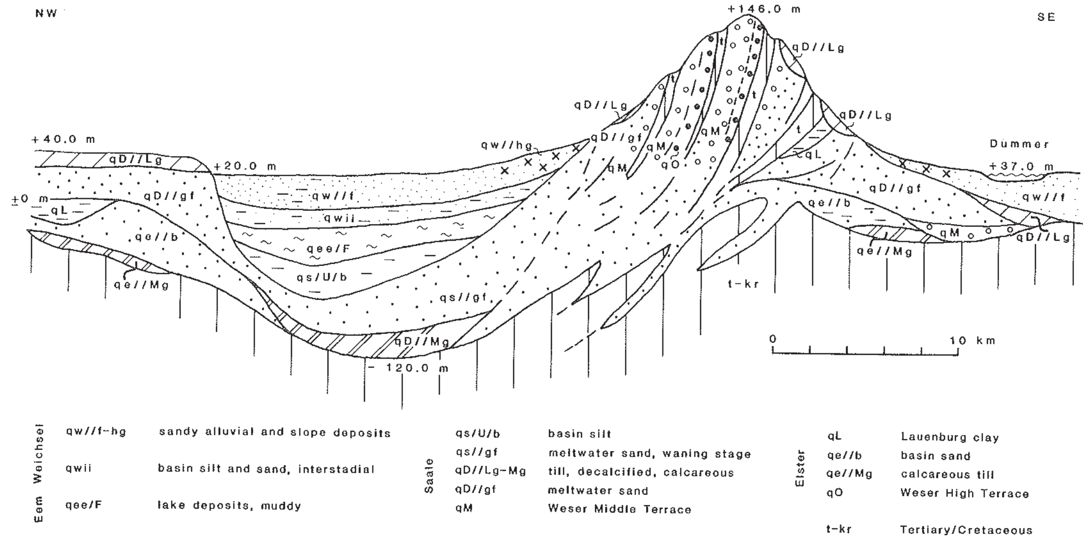


Figure 6.35 Quakenbruck glacial basin
Schematic profile through the Quackenbrück Basin and the Dammer Berge. As the base of the Eemian deposits is relatively low, this basin was probably a deep lake during the whole deglaciation (from: Meyer, 1987 - figure 7).

bedrock (LBEG, 2006). Its infilling consists of glaciofluvial sands and gravels (Van den Berg & Den Otter, 1993), most likely these deposits were transported to the basin as bedload by the deglaciation rivers from the Münsterland Embayment and the Weserbergland. Between Oldenzaal and Ootmarsum a depression in the ice-pushed ridge occurs. Here, the The Nordhorn Basin is connected to the Wilsum basin. To the north the Nordhorn Basin was also most likely connected with the basins near Lingen which probably mainly drained to the north trough the Hunze valley.

7. Classical glaciation phase models

During the last decades, several phase models have been developed. Most of these classical phase models were based on specific themes or biased to subregions (figure 7.1 and table 7.1). In this chapter the classical phase models and their lines of reasoning are outlined. Most of the elements these models refer to are described in chapter 6. As described in chapter 2.4.4 most of these classical phase models were also included in the GIS to compare them more easily, to show which elements were taken into account and which were not and to demonstrate the white spots in the classical phase models. Finally, their useful elements and the elements conflicting with other data are discussed. The different phase models are arranged by the features they are mainly based on.

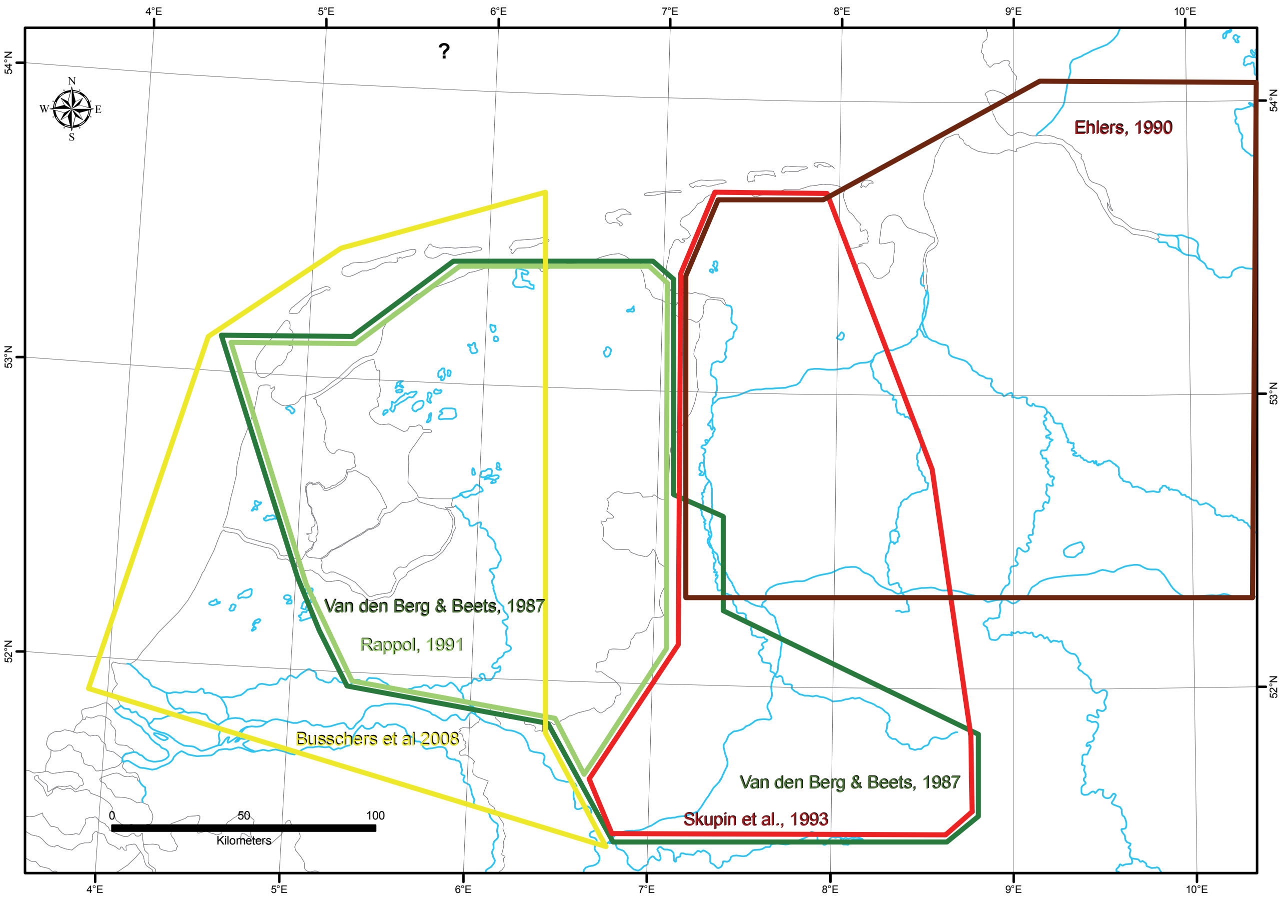


Figure 7.1
The research area of previous studies. The map shows the regional areas that were considered in the several classical phase models. Within these regions, the actually studied local sites usually comprise a much smaller region.

		Maarleveld	Ter Wee	vd B &B	Ehlers	Rappol	Skupin et al	Busschers	New phase model
Till stratigraphy	*Erratics*				x	-	x		x
	Orientational structures				x	x	x		x
Till sheet morphology	*Flutes, drumlins*			x	/	x	/		x
Push moraines	*Strike-dip/pushing direction*	x	x	x		x			x
	Strat (overridden, readvance)		x	x		x	/		x
Proglacial features	*Sandurs*			x				x	x
	Kames							/	x
	Ice-marginal rivers		x					x	x
	Ice-marginal lakes							x	x
Deglacial features	*Intramarginal rivers*			x				x	x
	Intramarginal lakes							x	x
Other	*Eskers*			x			x		x
Substrate				x					x

Table 7.1
This table shows which of the different aspects of the glacial, proglacial and deglacial situation were considered into several classical phase models.

7.1 Phase models based on geomorphology

7.1.1 Central Netherlands – Maarleveld

Maarleveld (1953; 1981) developed a regional phase model for the ice-pushed ridges in the central Netherlands based on morphology, intersecting relations and measurements of strikes and dips of thrust sheets in the ice-pushed ridges. His model used the following assumptions:

a. In the presence of relief, land ice does not spread in single ice front, but in lobes.
b. The strike of the thrusts and nappes in the ice-pushed ridges is perpendicular to the direction of pushing
c. The shape of a basin is a reflection of the last occupation by ice.
d. The age of the ice-pushed ridges within ice pushed ridge complexes diminishes towards the glacial basin. These sequences of ice-pushed ridges represent waning stages of land ice, i.e: the ice-pushed ridges were formed in stages and that each was formed during a re-advance within the general recession of the ice margin.

The principles were used to deduce the relative chronology of the formation of the several ice-pushed ridges was deduced. For example: the Gooi ice-pushed ridges and the Den Dolder-Amersfoort ice-pushed ridge were formed after the more southern parts of the Utrecht Ridge (for placenames and the ridges see figure 7.2). Within the Gooi area the Hollandse Rading-Hilversum ridge is older than the Laren – Huizen ridge. The configuration of the ice-pushed ridges of the Gelderse Valley was explained as follows: the Soesterberg ridge, the Kootwijk-Putten ridge and the Oud Reemst ridge were of the same age, both younger than the Austerlitz, Amerongen, Garderen and Ede ridges (all these ridges were presumably formed by waning ice lobes).
Near Nijmegen, the ice-pushed ridge is relatively wide, indicating two ice pushed ridges formed in two

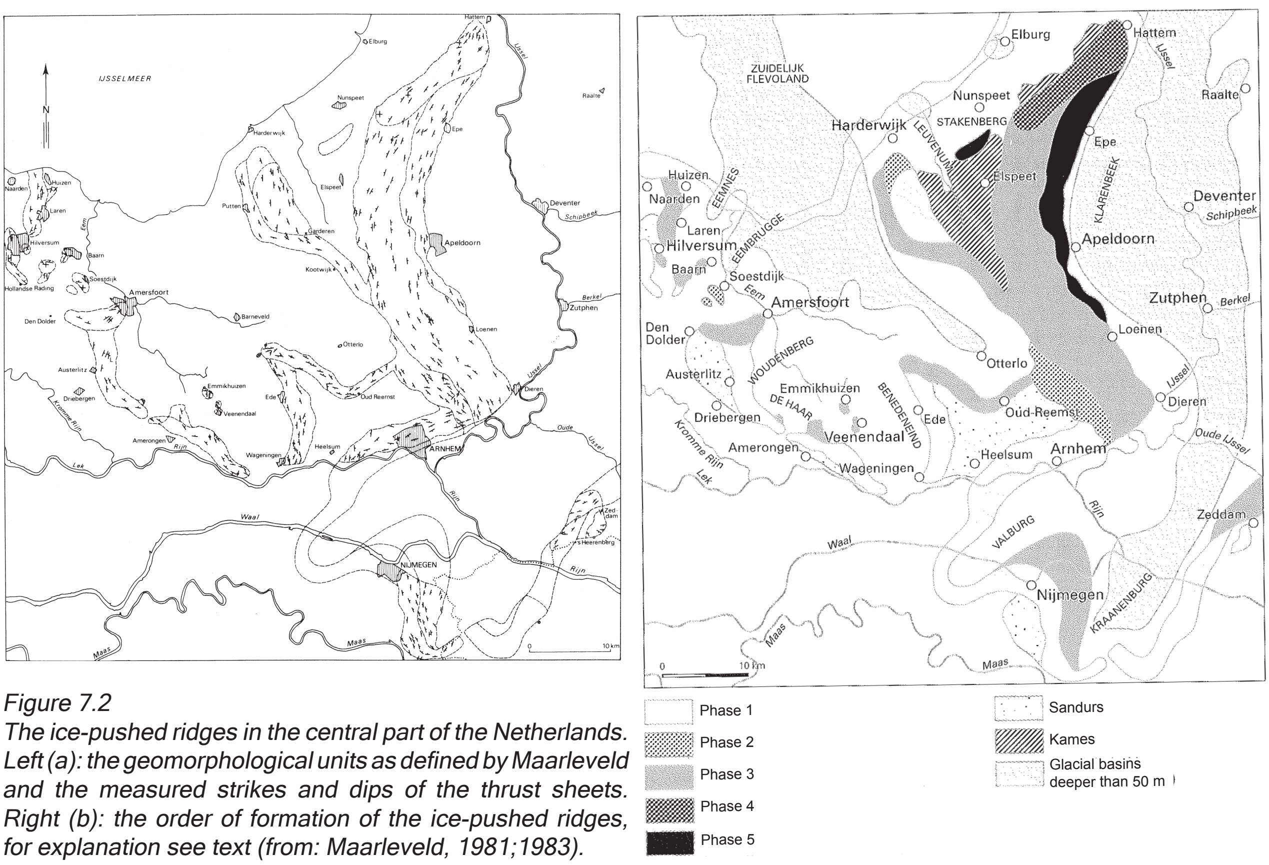


Figure 7.2
The ice-pushed ridges in the central part of the Netherlands. Left (a): the geomorphological units as defined by Maarleveld and the measured strikes and dips of the thrust sheets. Right (b): the order of formation of the ice-pushed ridges, for explanation see text (from: Maarleveld, 1981;1983).

phases. Because the the Valburg-Kranenburg lobe is the most far away from the glacial basin (IJssel basins) it is thought to be the oldest lobe in this region. It was partly reworked by ice lobes pushing from the west forming the Eastern Veluwe ridges. The slight curving to the east of the Eastern Veluwe ice-pushed ridge in the northern part was explained by a younger pushing phase from the northwest. This is confirmed by deflection of the strikes in the Woldberg.

Discussion
The assumptions a b and c are confirmed by evidence from the study of Maarleveld and by later studies. A lobated ice front was for example also suggested by Rappol and Van den Berg & Beets. The assumption that the ridge formed parallel to the ice pressure was confirmed by Van der Wateren (1995). The idea that the ridges in the Central Netherlands represent waning ice lobes was applied by Ter Wee to the whole Netherlands. For the central part of the Netherlands the idea of waning ice lobes was adapted in the Gooi region (Ruegg, 1977; Ruegg & Koopman, 2010) and the Amersfoort ridge (Van Balen, 2007). It was however rejected for the region of Arnhem by Bakker (2006). He concludes that the Arnhem ridge was formed at least partly during the formation of the Eastern Veluwe ice-pushed ridge and not entirely before. Implying that the ice lobe in the Gelderse Poort occurred later that the ice lobe that formed the Eastern Veluwe.
Another problem is that some of the units defined by Maarleveld, in reality do not seem to exist. Bakker (2006) concluded that the Eastern Veluwe is the result of only one major pushing event expressed in a distinct number of glaciotectonic styles (figure 6.17). Layers that dipped in an opposite direction on the distal part of the ice-pushed ridge can be attributed rather to folding of the layers than to different pushing phases. This is confirmed by the AHN (figure 6.18), here just one ice-pushed ridge can be seen. Only the Woldberg in the north can be considered a separate unit, it was formed by ice lobes flowing west and east of the Veluwe. In general, it can be concluded that the ridges formed during a advance of the ice lobes. At some locations (Gooi area) minor local readvances occurred. In the new phase model presented in this study, these ice front oscillations are considered to be too local to correlate to regional phases, and are therefore not considered to represent more than one phase. For the same reason, the elements of this phase model are not included in the GIS database.

7.1.2 Ter Wee
Ter Wee (1962) developed a model based on the distribution of ice-pushed ridges and river valleys. The ice-pushed ridges in the central part of the Netherlands were formed during two phases of an advancing ice front (contra Maarleveld). The ice-pushed ridges in the Eastern and Northern Netherlands were formed during a readvance of the ice front in the last three phases (D,E,F). The Rhine, Vecht and Hunze valley were interpreted as ice-marginal rivers close to the ice margin (Ter Wee, 1962; 1966; 1979). Jelgersma & Breeuwer (1975) refined the model and also interpreted the ridge complexes as recession lines of the ice front.

Discussion
Although this model explains the distribution and shape of all the ice-pushed ridges in the Netherlands, the main assumption of the model is proved to be invalid (Zonneveld, 1975). The ice-pushed ridges in the Northern Netherlands and some in the Eastern part are covered with tills and some of them have a drumlinized shape (chapter 6.2). When the ice-pushed ridges north of the maximal extent are younger, one would expect that older meltwater deposits and tills would be incorporated in the ice-pushed ridge. This sedimentological evidence was never found for most of these moraines (chapter 6.2 - Van der Wateren, 1995). All this strongly indicates overriding and thus an older age of the northern ice-pushed ridges. In Germany the same pattern can be seen in the Rehburg ice-pushed ridges, that were also overridden (chapter 6.2). The idea that the Vecht and Hunze valleys are readvance pradolinas is very unlikely because no large scale meltwater deposits were found of sandurs along the ice-pushed ridges

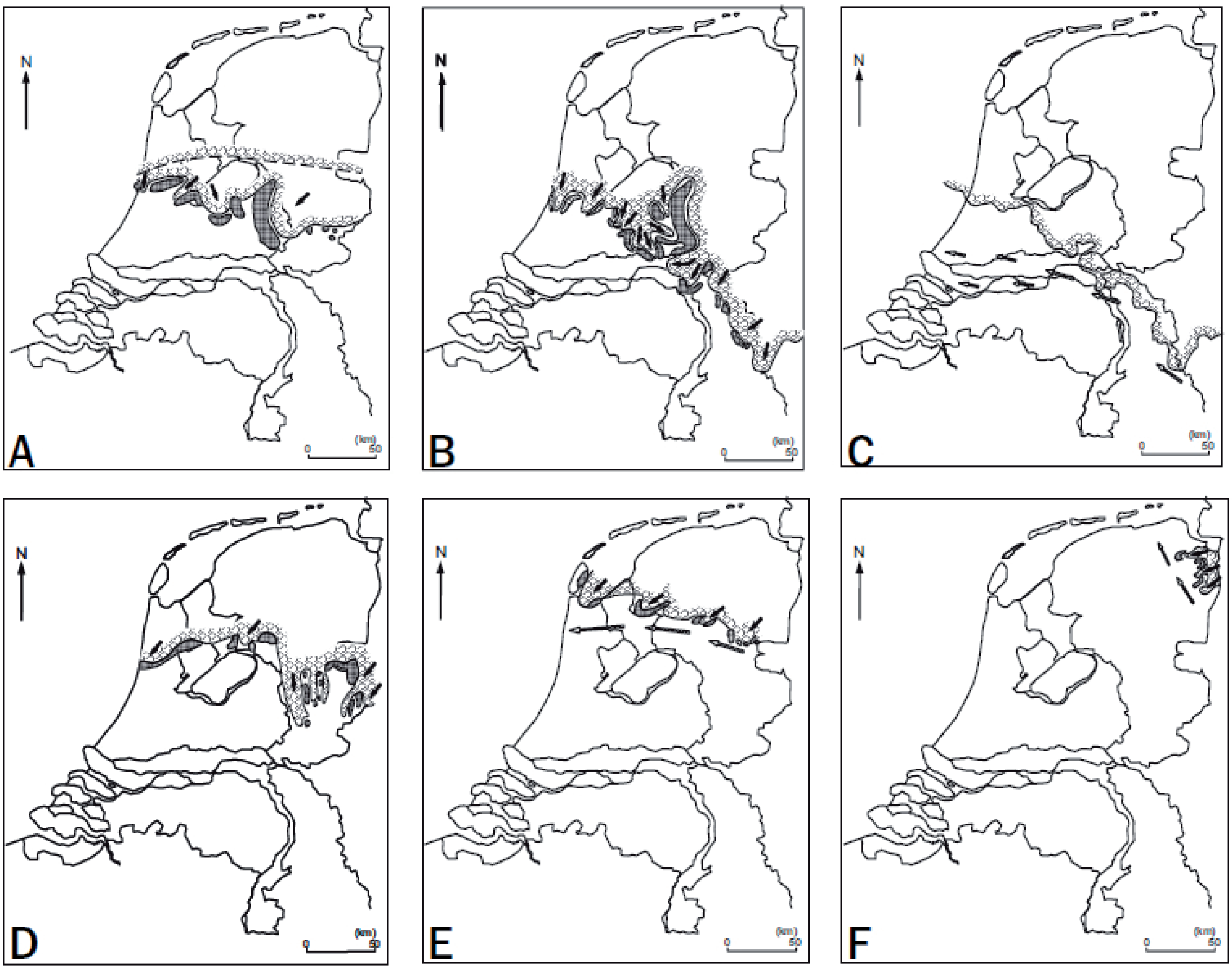


Figure 7.3
Phase model of Ter Wee, for explanation see text (from: Jelgersma & Breeuwer, 1975; after: Ter Wee, 1962).

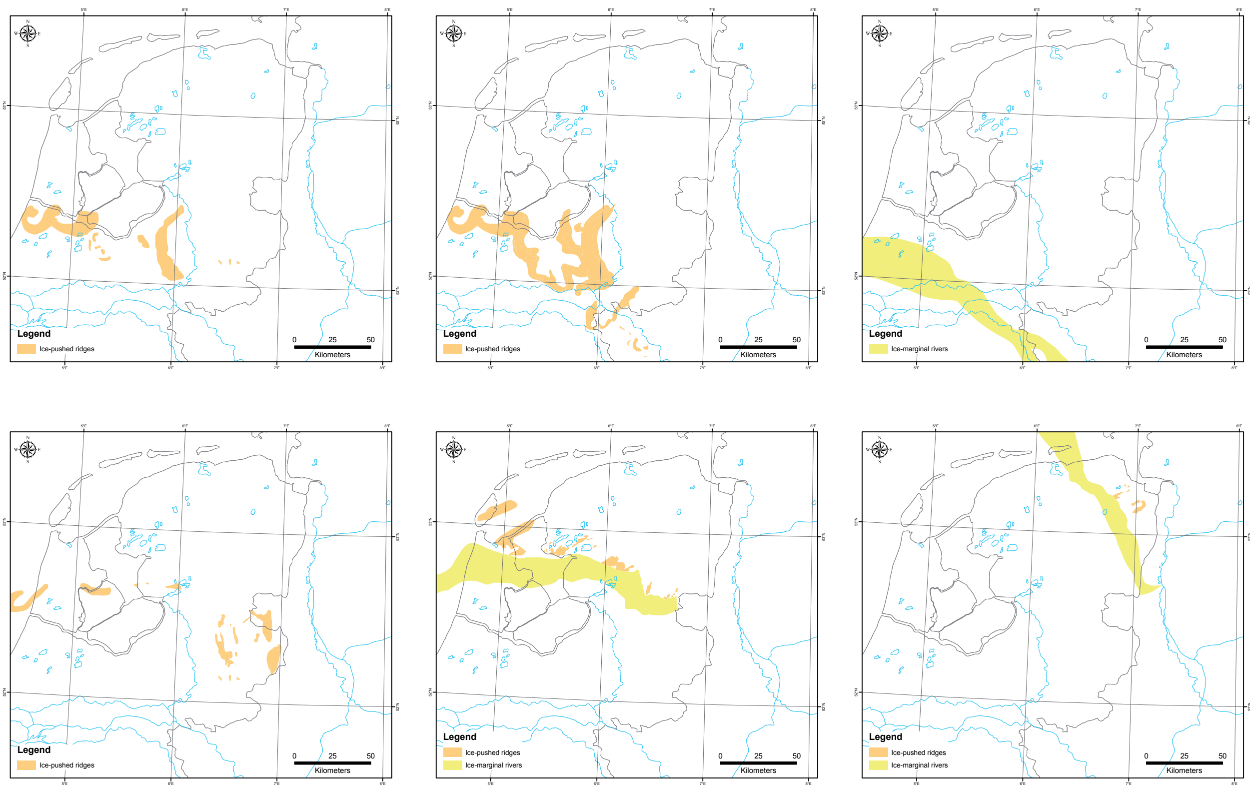


Figure 7.4
The phase model of Ter Wee reproduced from the GIS.

into the valley (Van den Berg & Beets, 1987). Other arguments can be derived from the fact that the valleys are deeply incised into the subsoil (see chapter 6.5.2).
Another complication of the model is the discussion about the 'ice-pushed ridges' formed in phase F in northeastern Groningen. Most likely, these are just erosional remnants of a till sheet instead of ice-pushed ridges (chapter 6.2.4). Besides, Ter Wee used exceptionally long and narrow glaciers to explain the ice-pushed ridges in the eastern Netherlands, which are now known to be unrealistic (Bakker, 2006). Also, the explicit morphological feature of the Hondsrug (chapter 6.1.3) and the distribution of erratics on it were not explained in this model.

7.1.3 Van den Berg & Beets

By the late eighties the recessional model of Ter Wee was considered to be unrealistic and a new model was needed. Van den Berg & Beets (1987) made this new model considering the following premises:

a) One basal till was found, suggesting only one ice cover,
b) The NE-SW of the Drenthe plateau and the NNW-SSE lineations of the Hondsrug have a glacial origin (contra Ter Wee, 1979 - chapter 6.1.3),
c) The ice-pushed ridges have been formed during an advance of the ice front,
d) The ice-pushed ridges in the northern and some in the eastern part of the Netherlands were overridden by the ice front (for discussion see chapter 6.2 and 7.1.2).

The sparse existence of ice-pushed ridges in the north and northeastern part of the study area, marks rapid extension of the ice front over fine grained, relative impermeable substrate (chapter 3.7). This advance came from the NE and slowed down over coarser substrate, which acted as a kind of ice trap (chapter 5.3). This created an even more strongly lobed ice front which resulted in the formation of horseshoe shaped ice-pushed ridges.
Van den Berg & Beets (1987) correlated the advance into the central Netherlands to the Rehburg line in Germany. In the eastern Netherlands, near Ootmarsum the ice-pushed ridges from this phase were overridden, this is due to the fact that finer Tertiary sediments are present in the subsoil in this part of the Netherlands. This allowed for acceleration of the ice draining from the north along a NNW-SSE flow over the Hondsrug area (figure 7.5 and 7.6). The ice mass surrounding the flowing ice was cut off its source and became dead ice. On Antarctica similar ice flows in a dead ice field were recognized (Hughes et al., 1985). The flow advanced into the Münsterland Embayment and caused the Hondsrug complex to form. During the deglaciation the Vecht and Hunze valley were formed that probably drained glacial basins in Germany.

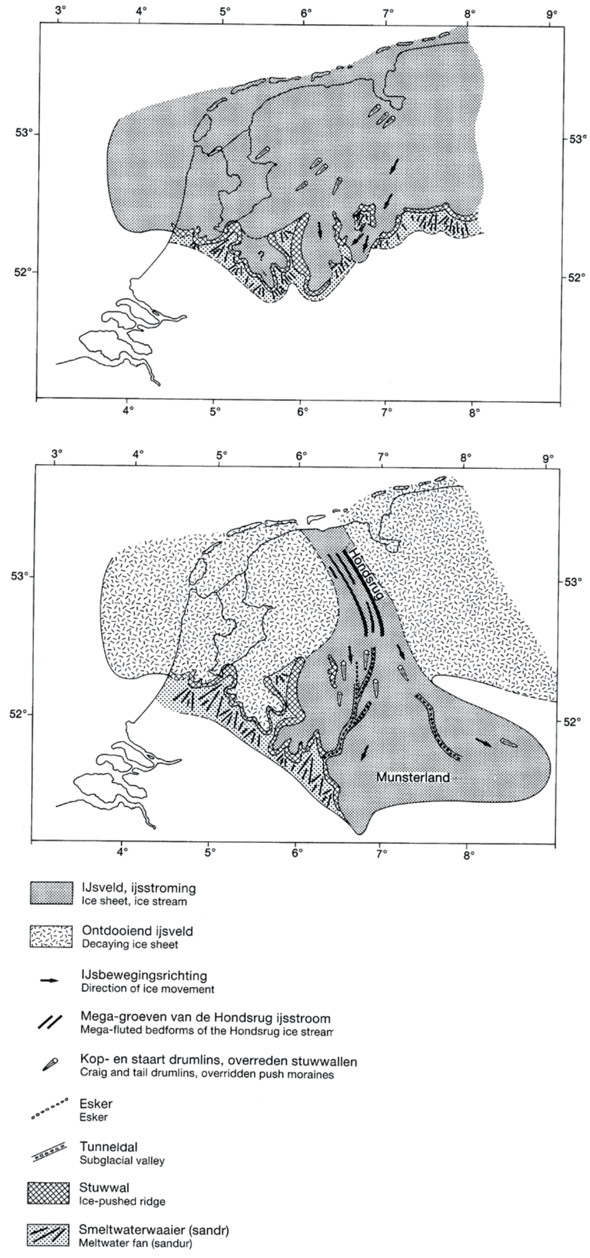


Figure 7.5
Phase model of Van der Berg & Beets, for explanation see text (from: Van den Berg & Beets, 1987).

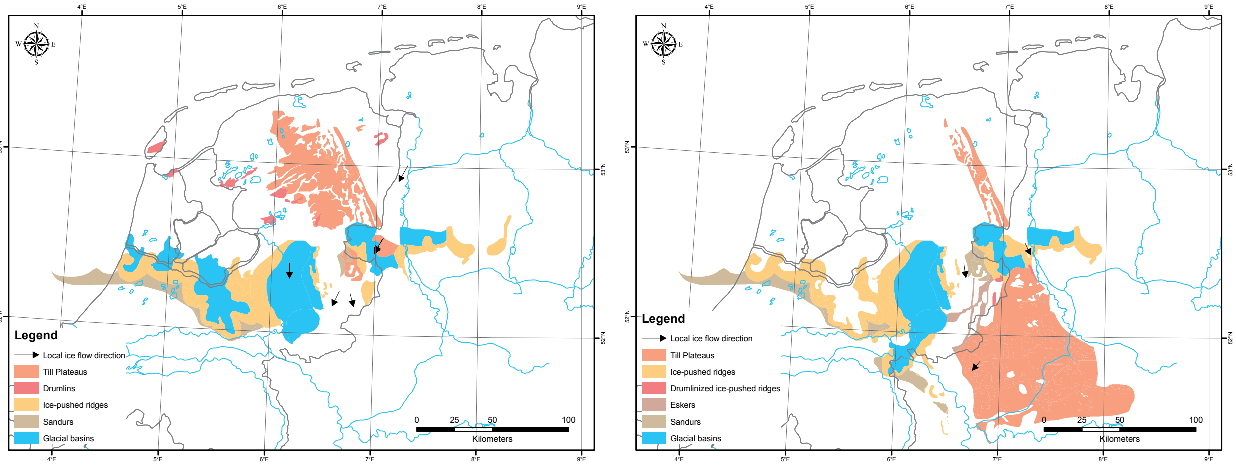


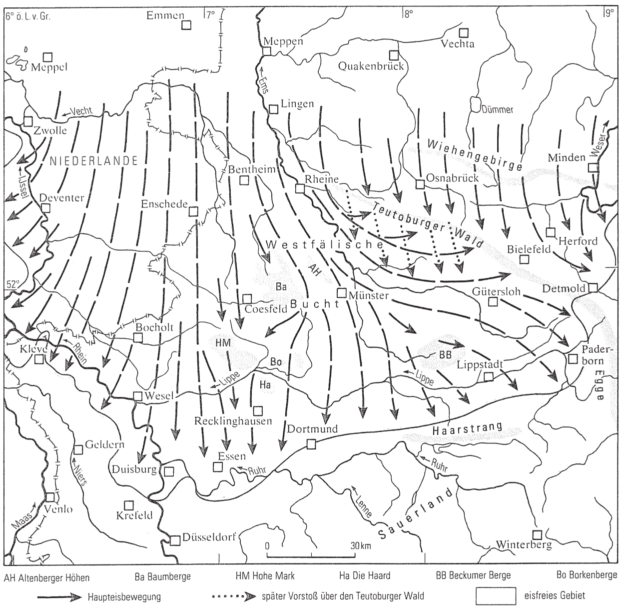
Figure 7.6
Comparison of the phase model with the mapped glacial features from the GIS in this research.

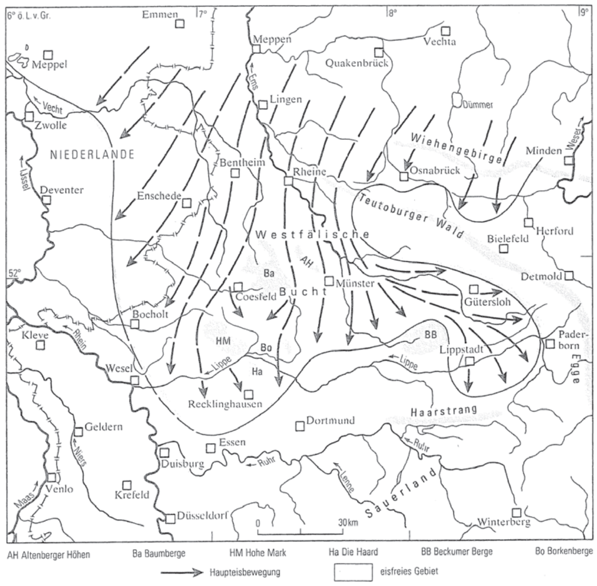
Discussion
This model nicely includes most of the geomorphological glacial features and their chronological relations. Besides, the order of the formation of the ice-pushed ridges in the middle and northern part of the Netherlands was displayed in a correct order, in accordance with the evidence of the overridden ice-pushed ridges.
The relation between the coarse materials in the subsoil and the formation of ice-pushed ridges seems to be a likely mechanism, this is confirmed by the ice flow pattern of phase 3 in the new phase model. Probably, more mechanisms are involved in the formation of the ice-pushed ridges (chapter 5.3).
This was the first model to link the Hondsrug formation with features in adjacent areas. The intrusion of an ice stream into the Münsterland Embayment was later confirmed by Skupin et al., (1993 - chapter 7.2.1). The overriding of the Twente and Itterbeck-Uelsen ice-pushed ridges by this ice flow was confirmed by the drumlinoid orientation of the overridden ice-pushed ridges in Twente (chapter 6.2) and the till stratigraphy (chapter 6.1). The initiation of this ice stream is still debated (see discussion chapter 7.3.2 and chapter 8.4). The deflection of the ice flow when the substrate allows faster ice flow, as suggested by Van den Berg & Beets, does still not explain the change in direction of the ice flow.
The extension of the Hondsrug flow in Drenthe is located too far to the west in this model, as can be concluded from evidence of erratics and geomorphology (compare figure 7.5 and 7.6). The erratic assemblage (chapter 6.1) and geomorphological evidence from the IJssel basin and western Twente contradict the presence of this ice stream in this region. Therefore, the Hondsrug ice stream is reproduced in the new phase model with some modifications.

7.2 Phase models based on till stratigraphy and erratic distribution

In the eighties the composition of tills and especially the erratic content was considered to be a key to the reconstruction of the ice sheet in the Netherlands. Based on this, till stratigraphy was developed (chapter 6.1) which was subsequently linked to different glaciation phases. This method was also applied in Germany, where it became a major key to reconstruct glaciations.

7.2.1 Skupin et al. 1993; Speetzen & Zandstra 2009
Skupin et al. (1993) distinguished four glaciation phases in this area, three of them could be clearly seen. They based this model on till research, glacier striae and erratic distribution. Liedtke (1981) and Skupin et al. (1993) investigated the glaciation in the Westfälische Bucht and the Münsterland Embayment. Seraphim (1979) and Liedtke (1981) suggested that the ice prograded in the valley of the river Ems and spread out in the Münsterland Embayment. The Ems valley marks the pathway along which the ice





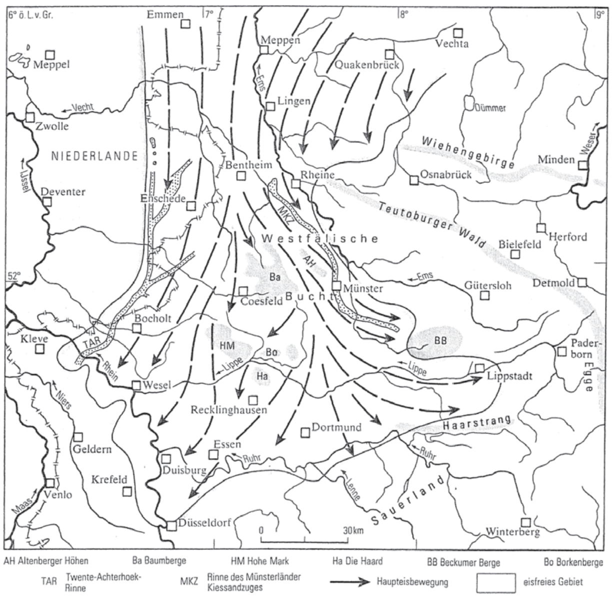


Figure 7.7
The phase model of Skupin et al., 1993. The situation in Lower Saxony was not yet incorporated in this figure.

was least hindered by the bedrock obstacles of the Weserbergland (chapter 3.7). At a smaller scale the Baumberge and the Beckermer Berge also caused the ice to deflect. This model was extended to Lower Saxony by Speetzen & Zandstra (2009). Below, these two models are integrated into one single model.

The first advance is the maximal advance and is called the 'main Drenthe glaciation'. It came from the north in Lower Saxony and it was deflected into the Münsterland Embayment (figure 7.7a). It formed the striking NNW-SSE morphology of the Syke Geest and the Rehburg ice-pushed ridges, which were subsequently overridden when the ice entered the Weserbergland and the Münsterland Embayment. A minor flow overrode the lower areas of the Teutenburger Wald and the Wiehengebirge bedrock area. This ice flow is associated with the Heerenveen till group containing mainly southern Swedish erratics to the area (area 7; III, Smaland) and some flint from Denmark.

The second advance is marked by erratics from Smaland (area 7, III) and Dalarna (area 6; II) due to a small change in source area. It was probably caused by a weaker and slower flow. It advanced over Ostfriesland and the Hümmling in a SW direction overprinting the structures of the first advance (figure 7.7b). It prograded to the Netherlands and the Münsterland Embayment but not as far as the first advance. It is associated with the flint rich till type within the Heerenveen till group. The transition between the first and second advance is thought to have happened more or less continuously.

During the third advance erratics from the Aland isles (area 2;I), the Baltic sea and Denmark were deposited. Tills with east Baltic components (the flint rich till type of the Assen till group) are typical for this advance. From this assemblage, Skupin and co-workers concluded that the flow path and source area of the ice must have changed (figure 7.7c). In the research area the ice stagnated and became dead ice. A 40-50 km wide ice stream followed a NNW-SSE direction, forming the Hondsrug reaching the Hümmling. This is evident from the erratic assemblage that is associated with this advance and from fabrics on the Hümmling (chapter 6.2). It continued south between Enschede and Rheine into the Münsterland Embayment where it reached somewhat further south that the second advance (chapter 6.1.7). The edges of the ice flow here are marked by the gravel ridge in the Münsterland Embayment and the tunnel valley in the Achterhoek-Twente. According to Speetzen & Zandstra (2009), ice flow could follow the Hondsrug through a large inherited valley. This valley probably formed during the Elsterian or it resembles the Saalian Weser/Ems valley.

The fourth phase left a smaller amount of till (Assen till group) that is marked by a remarkable absence of flint in the east Baltic tills and erratics, it was only described for the Münsterland Embayment.

Discussion

This glaciation model is based on a mono-thematic approach, till stratigraphy and the distribution of erratics are considered the key indicators to reconstruct the phasing. To complete the phase model, the overridden Rehburg ice-pushed ridges and the morphology of glacial lineations (deduced from drainage patterns of local rivers) were also taken into account.

The model was not included in the GIS, it is not compatible with the other geomorphological features included in the model. The monothematic approach is the major drawback of this model because it overestimates the importance of till types and erratics. There are two objections against a monothematic approach: (I) other important features and mechanisms are overlooked; (II) the relation between till stratigraphy, erratic assemblages and ice streams is too complicated to use it for reconstructions.

I) The importance of a multi-thematic approach is outlined in chapter 2. In the new phase model, a phase is defined as a distinctive ice flow at the ice margin that left significant geomorphological evidence. From this multi-thematic view, it would be inappropriate to mark a slight change in origin of the ice stream as a new phase (Skupin - phase 2) as this phase does not leave significant geomorphological evidence (only lithostratigraphically) in the study area other than phase 1. On the other hand, from this geomorphological view, it would be inappropriate to mark Skupin phase 1 as one single phase. Because the flutes of the Syke plateau and their truncations and the formation and overriding of the Rehburg phase clearly mark distinct geomorphological events. In the new phase model they are incorporated as such (transition phase 1 to phase 2).

II) The assumption that every till type was deposited by a different ice stream from a different ice flow is fundamentally wrong (cf. Rappol et al., 1991- see chapter 5.1.3). One single ice stream can carry different erratic assemblages, whereas one single assemblage does not necessarily indicate one single ice stream. This is because the factors involved in till composition are rather complicated. It is for example possible that one ice stream can deposit patches of various erratic assemblages (e.g. flint rich and flint poor - chapter 4.5) till as demonstrated by Rappol (1991a). This insight automatically makes the existence of phase 4 less likely because the extra flints, that it is based on, could easily have been incorporated within the same ice flow.

The third advance in the Skupin model, is a very distinct phase, both in terms of glacial dynamics, till stratigraphy and geomorphology. It is comparable with the Hondsrug ice stream of Van den Berg &

Beets and Rappol. The Skupin model gives the best insight of the distribution of this ice stream in the Münsterland Embayment, therefore this element is incorporated into the new phase model (chapter 8.4). The erratic assemblage of this phase marks the presence of an ice stream in a dead ice field.

7.3 Phase models based on till stratigraphy and fabric analysis

7.3.1 Ehlers 1983/1990

Ehlers (1983; 1990a;b) has linked glaciation phases recognized in NW Germany to the Netherlands, he distinguished three glaciation phases: The oldest, middle and young glaciation. As ice-pushed ridges do not always represent the outermost margins of the ice, till stratigraphy was preferred above of end moraine morphostratigraphy (Ehlers, 1983). Like the phase model of Skupin, this model is mainly focussed on till stratigraphy. Major difference is the focus on till fabrics, which represents local ice flow, next to erratic assemblages, that reflect source areas and ice-flow pathways.
Three major till units were distinguished using erratic assemblages (chapter 6.1) that served as the base for the glaciation model. Within the older glaciation phase three movement directions are distinguished based on fabric analyses (Ehlers, 1990a;b).

The Saalian glaciation, older phase
The advance of the Older Saalian Glaciation covered almost all of Lower Saxony, crossed the Münster Bight (Seraphim, 1980; Thome, 1980a), reached the Lower Rhine and left behind enormous pushed end moraines. The most prominent German ice-pushed ridges of the Older Saalian glaciation are those of the Rehburger end moraine. The occurrence of lodgement till in the foreland and on top of the ice-pushed ridges indicates overriding by ice (Meyer, 1980; 1983 – cf. the Dutch situation as outlined by Van den Berg & Beets, 1987). The older phase was again split up in three distinct subphases, based on the deduced ice flow direction and associated source area of erratics (figure 7.8).

North sea ice - The direction from the WNW (North Sea direction) formed the Hondsrug lineations and reached the Hümmling (Schröder, 1978) and the area around Bremen and the Syke Geest (Ehlers, 1990a;b).

Radial ice – The main phase is the 'Radial ice' with ice from the NNE NE. Indications for this flow direction was found in the Netherlands by Rappol (1983; 1987), many measurements from the Hümmling area (Schröder, 1978), the landforms in the Oldenburg-East Frisian Geest and fluted

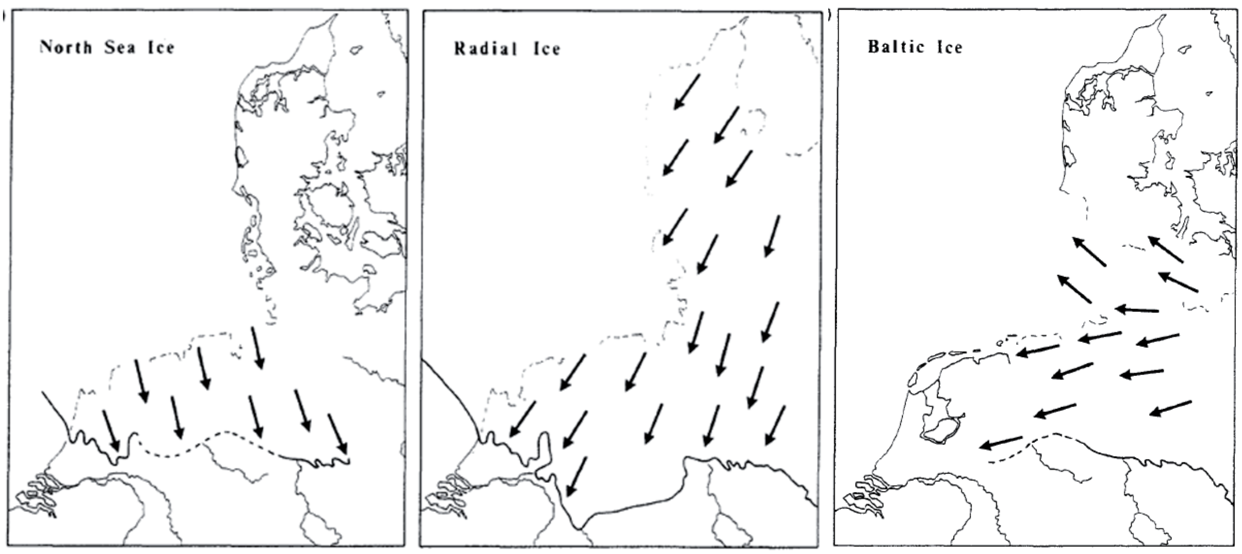
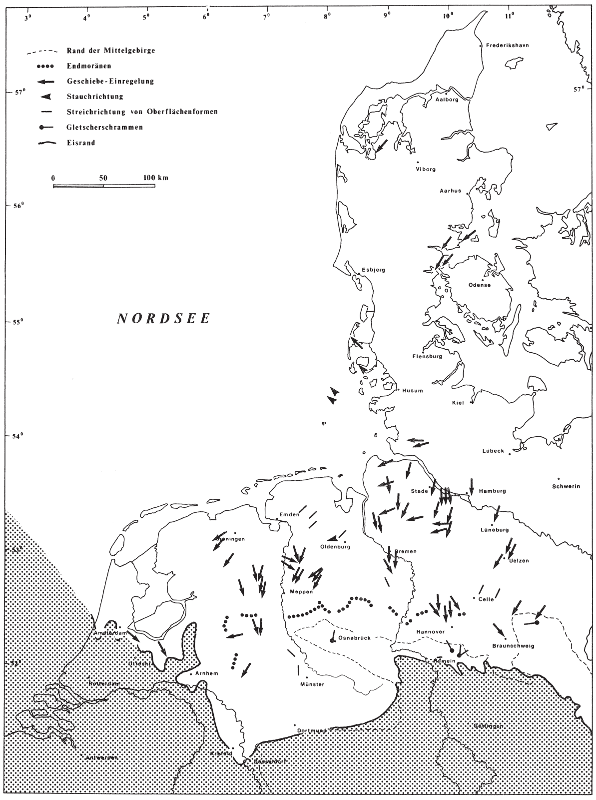
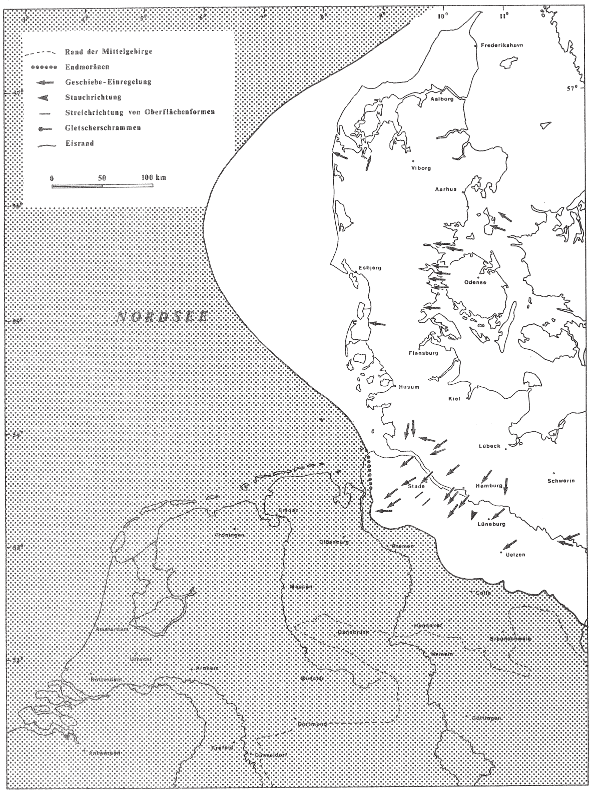


Figure 7.8a
The phase model of Ehlers. From: Ehlers, (1990ab)





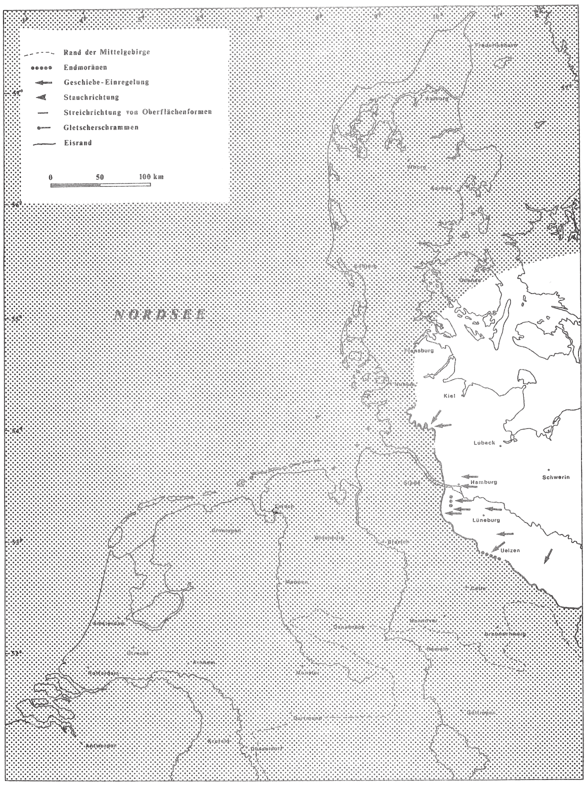


Figure 7.8b
The phase model of Ehlers: top left : the older Drenthe phase, top right: middle Drenthe phase, bottom right: younger Drenthe phase. From: Ehlers, (1990ab)

morphology in the Netherlands (chapter 6.1). According to Ehlers the radial ice advanced into the Münsterland Embayment instead of the NNW ice flow (North Sea ice) as suggested by Van den Berg & Beets (1987).

The Baltic ice- The third phase is called the Baltic ice advancing from the east from the Baltic and the East Baltic. Several NEE-SWW trending fabrics measured between the Elbe and Weser are considered to indicate this phase. Remarkable is the fact that Ehlers & Stephan (1983) mark these fabrics as the oldest fabrics in the area instead of the youngest.

The Saalian glaciation, middle phase

Around Hamburg a one meter thick sandy till was found representing this glaciation phase. It was separated from the Older Saalian till by meltwater deposits (chapter 6.1.6). Two principal flow directions were found, one from the north and one from the NE. Thick meltwater deposits in outwash fans on the Lüneburger Heide were formed in front of the ice front (Ehlers, 2005). This glaciation covered only a minor part of NW Germany, including the area around Hamburg. The meltwater drained southwards and then via the Aller-Weser ice-marginal valley towards the North Sea. At the end of this phase large dead ice masses are likely to have formed which were subsequently covered by outwash deposits from the next glaciation phase (Ehlers, 2005). This phase is called Drenthe-2 by Meyer (1987; 2005).

The Saalian glaciation, younger phase
A till containing a distinct erratic assemblage overlays meltwater deposits upon the till of the Middle phase (chapter 6.1.6). The distribution of this till suggests that the ice retreated further to the northeast (Ehlers, 2005). Till fabrics indicate an ice flow from an easterly direction. This glaciation phase is correlated to the Warthe glaciation by Meyer (1987; 2005). Another possibility is that both the middle and the younger phase are correlated to the Warthe (Ehlers, 1990). Like the Middle Saalian glaciation this glaciation only covered a small area of the research area. on.

Ehlers (1990a) suggests that changes in ice flow direction are caused by shifting the ice divide and by changes in the glacier dynamics (chapter 4.3.3). The ice divide shifted from west to east causing the ice flow in the research area to deflect from NNW (north sea ice, older Saalian glaciation) finally to the east during the younger Saalian glaciation phase.
Ehlers (1990a) also stated that the advance of the ice front must have been relatively fast. For the Weichselian it was estimated to be 100-150 m/yr. This rapid ice front progradation could also have occurred during the Saalian and it can at least be partially explained by the deformable bed (figure 4.B) under the ice (Ehlers 1990; Boulton & Jones, 1979). This mechanism causes the ice to thin towards the margins (Boulton et al., 1985).

Discussion
This research gives a very proper overview of the fabric analyses in the study area. Another advantage is the fact that both the Dutch and the German situation are considered. They are however linked in a wrong way. As with the model of Skupin, this model has a monothematical approach strongly focussing on till stratigraphy and fabric analyses. For the distinction between the older, middle and younger phase, this seem to be quite helpful. However, for the subdivision of the 'Old phase' this approach is less valuable and the subdivision of this phase conflicts strongly with other evidence. The first phase of the model of Ehlers is not included in the GIS, as it is not compatible with the other geomorphological features.
A major problem occurs when the model is matched to the geomorphological evidence (SRTM, figure 6.H). It clearly shows that the 'North Sea flow' should be split in two phases. The Hondsrug area and the Syke area may show the same fabrics and flute orientation, but from cross-cutting relations it becomes clear that the Syke flutes must be older than the Hondsrug. The generally trending N-S fabrics east of the Syke plateau and in Hamburg seem to match the oldest ice flow.
When the glaciodynamics are considered, it is very unlikely that a large distinct ice mass enters the same the area from three totally different directions. This means that all the ice from the previous advance must then (nearly) have disappeared, yielding large meltwater deposits. Another possibility is that the complete ice mass turned into dead ice and was totally pushed away. As no meltwater deposits have been found between the tills here (chapter 6.1) and proof lacks for the latter, the whole scenario is not very likely. Another possibility is that the ice flew rapidly and that this change in ice flow direction occurred gradually. This is not confirmed by the geomorphological evidence and fabrics as only three directions have been found, and intermediate directions seem to be lacking.
The model has strong focus on the area between the Weser and the Elbe, which is quite well described, and where two phases are distinguished. The complex glacial processes that took place west of this region are not well described and therefore the subphases within this phase do not match with (geomorphological) evidence from other studies. The tills from the Drente substage are incorporated in the ice-pushed ridges, which indicates glacial readvance (chapter 5.3.3, 6.2.6). The meltwater deposits between the Drente till and the tills from these phases (figure 6.J) also confirm the occurrence of two readvance phases, called the Middle and the Younger phase. Fabrics from the study of Ehlers and co-workers reveal that the first readvance came from the northeast and the second one from the east. This

can also be nicely connected to the shift of the major ice divide in Scandinavia as suggested by Ehlers, (1990 - chapter 4.3). The formation of the Wester-Aller ice-marginal river during the 'Middle phase' also matches geomorphological and stratigraphical evidence, although more of these valleys must have occurred (chapter 6.5).

7.3.2 Rappol

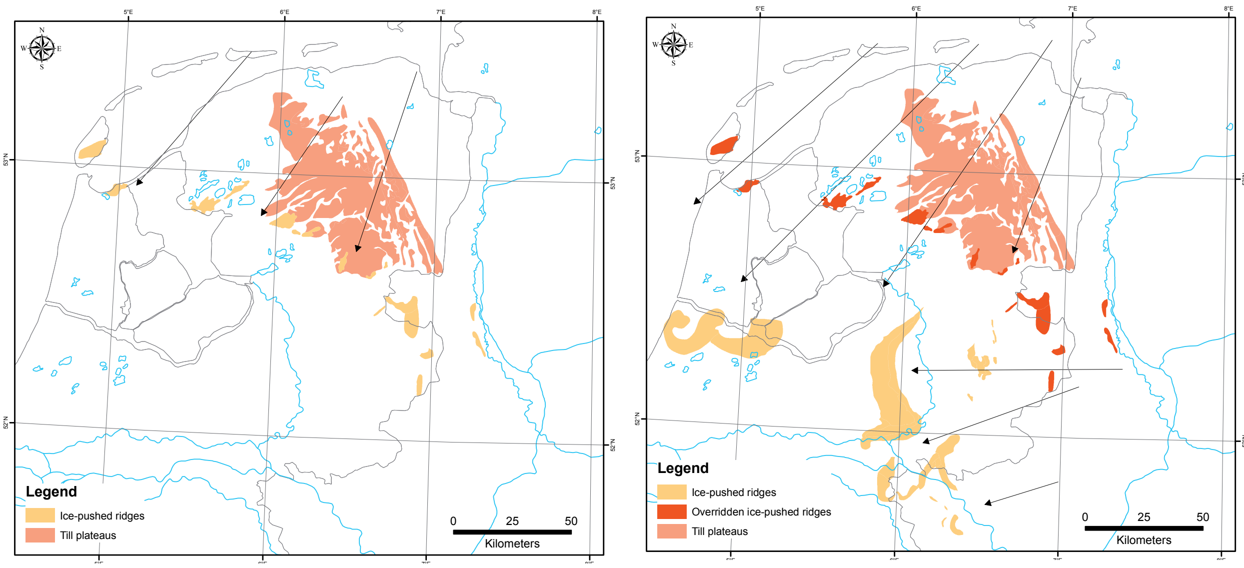
An integral approach of morphology, till stratigraphy and erratics was done by Rappol (1991a;b). The till stratigraphy and structural dynamic evidence from tills (Zandstra, 1983; Rappol, 1987) (chapter 6.1) was the most important input for this model. Ice-pushed ridges were assumed to have formed frontally in stead of laterally (chapter 5.3). The model distinguishes three phases (figure 7.9):

Phase 1: The ice advanced to the line Texel-Wieringen-Gaasterland-Steenwijk, it also covered the eastern part of Overijssel. From the thick till deposits (Voorst Till type) at the margins of this phase, Rappol concluded that the ice front must have been stagnant. About the direction of the ice is much uncertainty. However, till fabric measurements revealed a southward flow in the eastern part of Overijssel (Rappol et al., 1991) and a flow to the west on Wieringen (Rappol, 1991b). Possibly, the Emmen ice-pushed ridge under the Hondsrug and the sandurs of Itterbeck were formed during this phase (Rappol & Kluiving, 1992).

Phase 2: The ice-pushed ridges in the northern part of the Netherlands were formed and subsequently overridden (cf. Van den Berg & Beets, 1987). During this phase the ice movement was reactivated after stagnation during phase 1. This caused the ice thickness to increase and led to the pushing of the margin of phase 1. In the northern Netherlands the flow direction was to the SW (deduced from geomorphology and till fabrics - chapter 6.1.3) and the Heerenveen till group was deposited. Possibly, the Rehburg moraines were also formed during this phase. In the eastern part of the Netherlands the flow direction was to the west and the Markelo till type was deposited (chapter 6.1.4 - Kluiving et al., 1991). According to Rappol this westward flow also caused the formation of the eastern part of the Veluwe by frontal ice pushing.

Phase 3: Phase 3: In this phase the flow pattern changed radically. Distinct erratic assemblages and orientation of flutes indicate two ice streams from the NNW, one over the Hondsrug and into the Gelderse Vallei. An Eastern Baltic erratic assemblage (area 1 & 2) combined with the NNW-SSE oriented flutes and oriental structures in tills mark the Hondsrug ice-stream. The distinct erratic assemblage can be traced into the Münsterland Embayment, which led the authors conclude that the ice stream reached this region as well. Fabric analyses in the northwestern part of the Hümmling suggest that the Hondsrug ice-stream also affected this region (chapter 6.1.6. - Rappol & Kluiving, 1992). Judging from the NNW-SSE oriented flutes on the Syke Plateau a third stream may have been present here as well. The orientation of the ice-pushed ridges around the Gelderse Vallei suggest an ice lobe intruding from the NNW, besides a deviating erratic assemblage occurs here (from Area 4,5 – around Stockholm). From their orientation perpendicular to the presumed ice flow direction (Kluiving et al., 1991) claimed that the northern tip of the Veluwe (Woldberg) and the northern part of the Sallandse Heuvelrug the Gelderse Valley were also formed by the Gelderse Valley ice lobe. The areas between these flows were most likely not affected by this phase, as the fabrics indicate the flow from phase 2 without overprinting by phase 3. These areas are thought to have been covered by dead ice.

The rapid ice advance in phase 1 is explained by the fine sediments that would have been present in the Baltic area. When these fine sediments eroded the resistance of the subsoil increased causing a stagnation at the end of phase 1. The transition from phase 2 to phase 3 is explained to indicate that north of the Netherlands ice flow from Scandinavia met ice flow from Britain, which deflected the flow direction. Rappol (1991a) gives two possible explanations for the location of the ice streams of phase 3: they could have formed in relation to permafrost patterns or because of the composition of



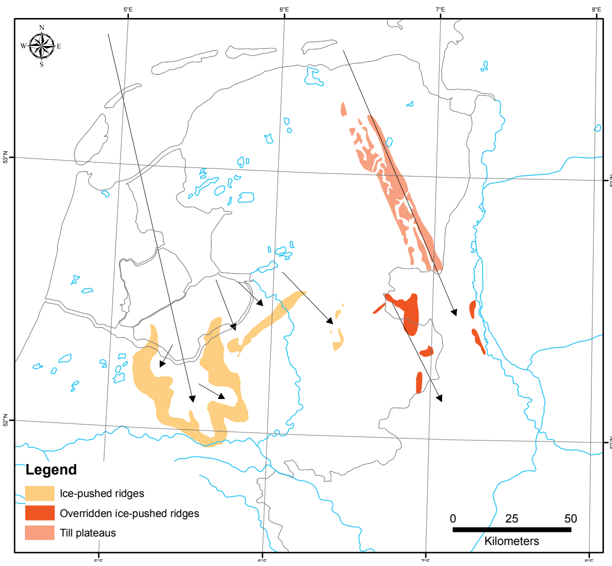


Figure 7.9
The three phases of Rappol, derived from the GIS (after: Rappol, 1990; Rappol et al., 1991).

the subsoil due to the location of preglacial rivers. From the lack of evidence of retreating phases it was concluded that the ice turned into dead ice after the glaciation.

Discussion

For the Northern Netherlands the sequence of events is quite well outlined by the Rappol model. The first flow direction from the NW spreading out to the SWW and SSW in the Netherlands can also be seen on elevation models and in Germany. Also, the younger age of the Hondsrug ice flow is very likely and can be traced into the Münsterland Embayment (cf. Van den Berg & Beets, 1987; Skupin et al 1993) and the Hümmling (confirmed by fabrics by Schröder, 1978 and erratics Speetzen & Zandstra 2009). The idea of the formation of dead ice (cf. Zonneveld, 1975 and Van den Berg & Beets, 1987) bodies matches the geomorphological evidence (truncation of the flutes – chapter 6.1)

Regarding the ice-pushed structures in the Northern Netherlands, the stagnation time of the ice is an issue of debate. The morphology can also be explained without the stagnation of the ice front. The ice-pushed ridges have formed due to the interaction of the ice sheet with the substrate and ice base hydrology rather than only by glacial dynamics (cf. Van der Wateren 1995).

Rappol assumed that the ice-pushed ridges were formed frontally with respect to the ice front. More recent studies reveal that both lateral and frontal pushing can take place (chapter 5.3.1). Therefore, frontal pushing of the Eastern Veluwe is now known to be an unlikely mechanism (Van der Wateren, 1995). The formation of the Woldberg and Sallandse Heuvelrug do not have to be formed by the

Geldersche Vallei ice stream (phase 3). In fact, the straight form of the Woldberg indicates that frontal pushing is not very likely. The presence of a very narrow ice lobe that crossed the IJssel Valley towards the Sallandse Heuvelrug is even more unlikely. These ice-pushed ridges were rather formed by an ice lobe splitting from the southward flowing large ice lobe in the IJssel valley (chapter 8.3).

The westward ice flow in the eastern Netherlands appears an unrealistic reconstruction. This was based on two fabric measurements (near Markelo and De Lutte), but is does not match the geomorphological evidence (figure 6.12 – push and drumlin like structures are oriented N-W instead of E-W). Moreover, in Germany no indication at all is present for this ice flow (figure 6.6 and 6.10). Most likely, the measured till fabric orientations are due to local ice flow directions caused by the deflection of ice around a ice-pushed ridge obstacle.

Another issue is the lacking of evidence for extensive glaciation of the North Sea (Joon et al., 1990) taking away the source for the flow that would have formed the Gelderse Vallei and the Hondsrug. Busschers et al. (2008) construct a large proglacial lake in the southern part of the North Sea. They state that in order to maintain the lake level, the British and the Fennoscandinavian ice sheets must have been in contact. This must have been firther north than nedded to support Rappols last phase. This does not necessarily mean that this collision actually caused the Hondsrug ice stream nor the Geldersce Valley ice stream. In fact it is very unlikely because such a collision would yield a much larger scale ice stream than the Hondsrug ice stream. A more likely mechanism for this ice flow is explained in chapter 8.4. No evidence of the Rappol-hypothesis of the northern extension the Gelderse Valley ice stream has been found. The differing erratic assemblage may be explained by the differential distribution of glacial debris from different levels in the main ice flow along the lobate ice front (see chapter 5.1.3 and discussion 7.2.1).

7.4 Phase models based on glaciofluvial sedimentology and stratigraphy

7.4.1 Klostermann, 1992

Klostermann (1992) developed a phase model that is based on distinct river courses, the preglacial relief and the morphology of the ice-pushed ridges that were formed in the Lower Rhine Embayment (LRE).

Onset

In 'Phase a' an EW trending, southward moving ice front was located near the 'Maarleveld line'. Between this line and the Wiehengebirge the Weser could drain towards the west. In the Netherlands de Rhine and Meuse were active. In the river beds the subsoil was relatively warm compared to the surrounding areas. 'Phase b' corresponds to the 'Rehburg phase' in which the ice front pushed up the large ice-pushed ridges north of the Wiehengebirge. In the Netherlands the ice front was located near Deventer-Almelo. The Weser and Rhine merged just south of this line.

During 'Phase c' these ice-pushed ridges were overridden (cf. Meyer 1980; 1983; Van den Berg & Beets, 1987) and the ice front could advance relatively quickly due to unfrozen subsoil of the former pradolina Weser. This river was blocked and a proglacial lake started to form. Favoured by the warmer river bed a glacial surge occurred in the IJssel basin forming the eastern Veluwe ice-pushed ridge and the Nijmegen-Kleve ice-pushed ridges (Kranenburger Lobus).

When the ice advanced further, the drainage from the Weser was totally blocked in 'Phase d' causing the extension of the large proglacial lake. The absence of the Weser caused the ice lobe to extent further southwards in 'Phase e' forming the Xantener Lobus and the Böninghardt outwash plain.

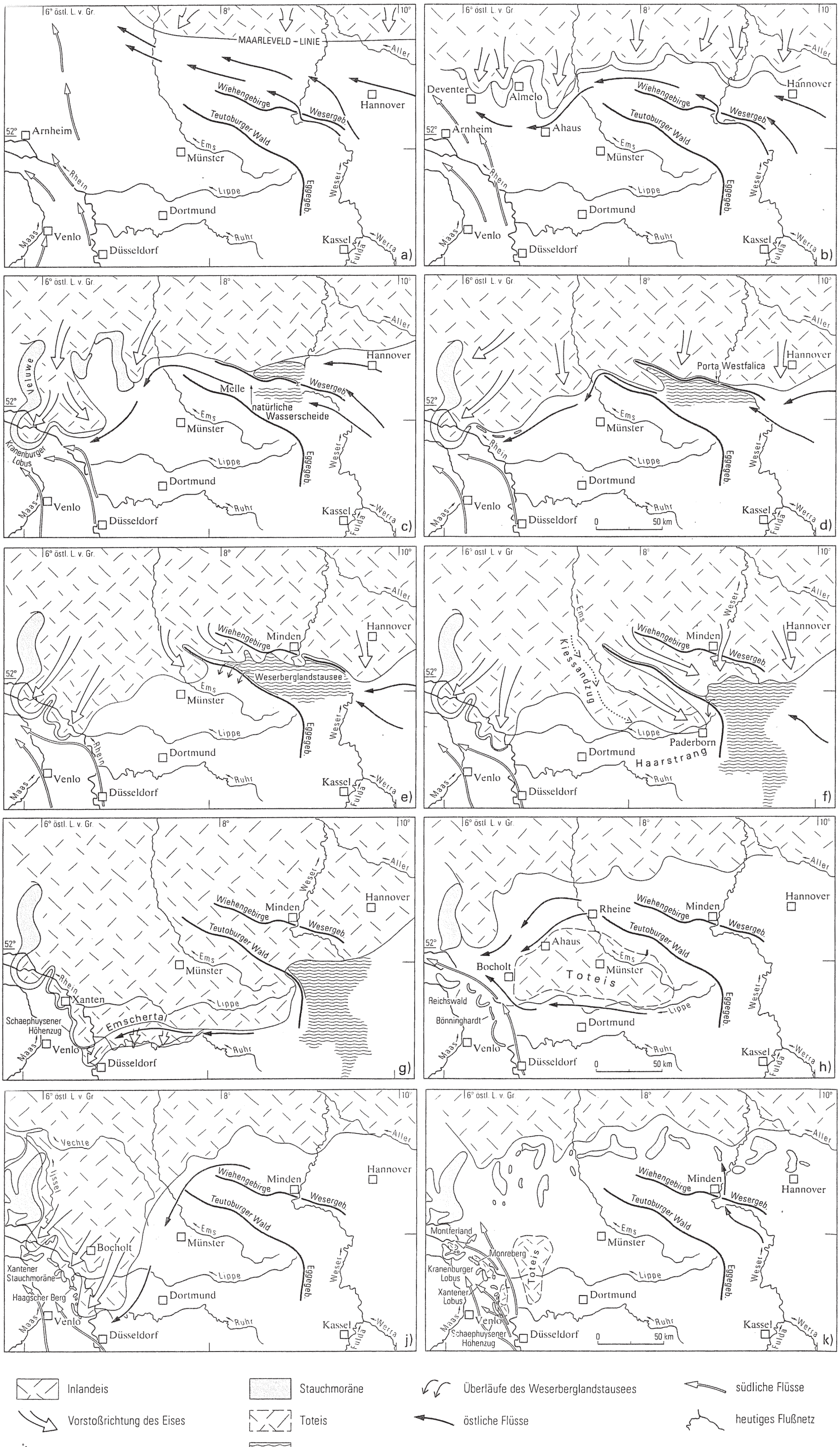
The extension of the ice between the Wiehengebirge and Teutoburgerwald caused the lake level to become so high that several overspills were formed in the Teutoburgerwald (Thome, 1983). This water

flow mobilised an ice stream south of the Teutoburgerwald into the Munsterbasin 'Phase f'.
The drainage of meltwater caused the ice front to expand even further into the Munsterbasin causing the maximal extension of the ice during 'Phase g'. The Schaepenhuysener ice-pushed ridge was formed during this phase. This phase is correlated to the Hamelner phase (Lüttig, 1959). In the LRE a minor oscillation caused the formation of several smaller ice-pushed ridges near Xanten on the intramarginal side of the larger ice-pushed ridges (Klostermann, 1989).

Deglaciation and readvance
In Phase h, some a large dead ice mass was formed in the Münsterbasin. The large proglacial Weser lake was able to drain large amounts of meltwater towards the west via Rheine, Ahaus and Bocholt. This meltwater eroded ice-pushed ridges located between the Vecht and Ems and east of the IJssel valley. Besides, the ice-pushed ridges in the LRE were partially eroded. This caused a new ice stream in this region in 'Phase j'. This ice stream deposited a distinct new till (Assen till group, Zandstra, 1993). The Xantener ice-pushed ridge and the ice-pushed ridges north of the Bönninghardt sandur were formed during this advance. Also the Montferland, Heyberg and Wolfsberg were pushed for the second time (Klostermann, 1985).

Discussion
This phase model was constructed mainly based on stratigraphical and morphological data
from the Lower Rhine Embayment, I.e. the regional model was constructed based on data from one single local area. This regional model was partly filled in by conceptual hypotheses rather than direct lithological evidence, e.g. the assumption that relative warm river beds cause surges.
In general, the order of events corresponds to other observation; progradation of the ice front, overriding of the ice-pushed ridges (cf. Van den Berg & Beets, 1987), deflection of the rivers (cf. Busschers et al., 2008), deglaciation. The large lobe in the IJssel basin was based on the assumption that latent heat from rivers could trigger ice streams. Probably, this played a role, but in this phase model it is overestimated.
Phase a corresponds with the preglacial configuration of the Weser (Maarleveld Line). It is quite likely that the Weser deflected towards the west in c/d, although this is not confirmed with any lithological evidence around the German-Dutch border. The diverging ice stream in the Münsterland Embayment as described by Skupin et al. (1993) is not incorporated in this model. The existence of the Lake Weser is confirmed by new data (e.g. Winsemann et al., 2010 – chapter 6.6.1). The initial drainage of the meltwater from Lake Weser probably did occur by overspillage in the Münsterland Embayment. When the ice advanced, this drainage mainly occurred south in the Münsterland Embayment (cf. Thome, 1983), and probably partially over the ice sheet. During deglaciation the ice probably retreated resulting in new overspilling into the Münsterland Embayment and the formation of Lake Munsterland (Winsemann pers. comm. – chapter 6.6.1, chapter 8.4).
The oscillations of the ice front in the LRE, deduced from the complexity of the ice-pushed ridges and the two tills that were found, may indeed be linked to the intrusion of the Hondsrug ice stream after the maximal extension of the ice. However, evidence for complete melting in the LRE and the presence of the Rhine is in between these phases is lacking. The model was not included in the GIS, because the model incorporates a lot of phases to form relatively small features in the LRE.

Figure 7.10 (next page)
Phase model of Klostermann, for explanation see text (from: Klostermann, 1992).



6° östl. L. v. Gr.
8°
10°
52°
MAARLEVELD – LINIE
Aller
Wiehengebirge
Wesergeb.
Teutoburger Wald
Eggegeb.
Hannover
Arnheim
Münster
Ems
Rhein
Lippe
Dortmund
Weser
Venlo
Maas
Düsseldorf
Ruhr
Kassel
Werra
Fulda
a)
Deventer
Almelo
Ahaus
b)
Veluwe
Kranenburger Lobus
Melle
natürliche Wasserscheide
c)
Porta Westfalica
0 50 km
d)
Minden
Weserberglandstausee
e)
Kiessandzug
Paderborn
Haarstrang
f)
Xanten
Schaephuysener Höhenzug
Emschertal
g)
Rheine
Bocholt
Toteis
Reichswald
Bönninghardt
h)
Vechte
Issel
Xantener Stauchmoräne
Haagscher Berg
j)
Montferland
Monreberg
Kranenburger Lobus
Xantener Lobus
Schaephuysener Höhenzug
k)
Inlandeis
Stauchmoräne
Überläufe des Weserberglandstausees
südliche Flüsse
Vorstoßrichtung des Eises
Toteis
östliche Flüsse
heutiges Flußnetz
subglaziale Schmelzwässer
Stausee

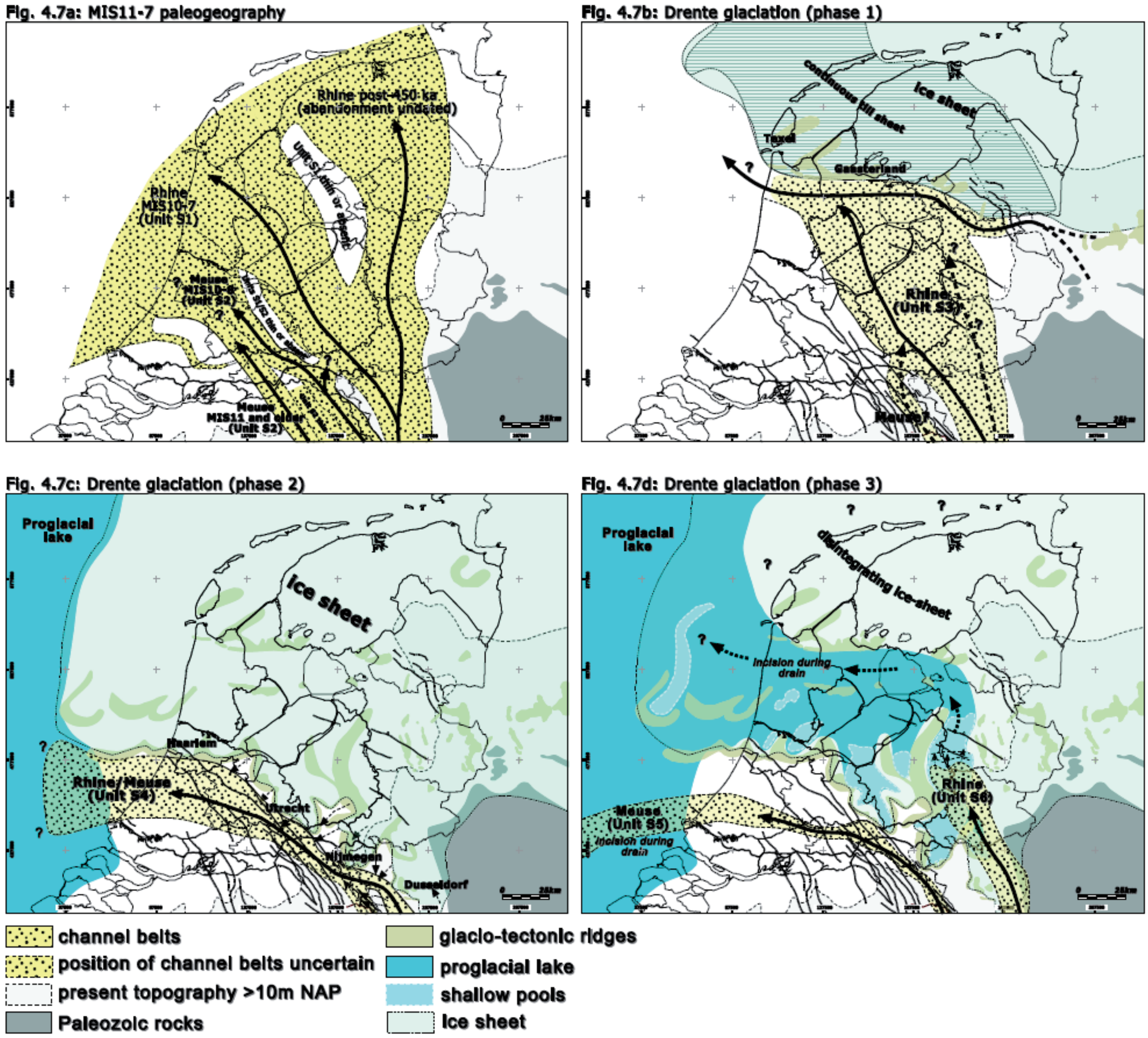


Figure 7.11
Phase model of Busschers, for explanation see text (from: Busschers, 2008).

7.4.2 Busschers, 2008

This phase model is based on sedimentological research on the Saalian Rhine and Meuse deposits it is focussed on the dynamics of the fluvial systems rather than on ice dynamics. Unlike most other phase models, this phase models also includes the preglacial and the deglaciation phases. Besides, it is the first model that provides absolute chronology from OSL dates. Therefore, integrating this model with other knowledge helps phase models integration.

Preglacial situation

This phase model is based on sedimentological research on the Saalian Rhine and Meuse deposits. It is focussed on the dynamics of the fluvial systems rather than on ice dynamics. Unlike most other phase models, this phase models also includes the preglacial and the deglaciation phases. Besides, it is the first model that provides absolute chronology from OSL dates. Therefore, integrating this model with other knowledge helps to integrate phase models.

Phase 1: During this phase an advance of the ice sheet into the Northern Netherlands up to the Gaasterland-Texel line. In the Central Netherlands deposition occurred of Rhine - Meuse (‘unit S3’) with an age of 168 +/-19 ka. Probably this system made contact with the proglacial drainage of the Northern Netherlands and ice-marginal drainage from Germany.

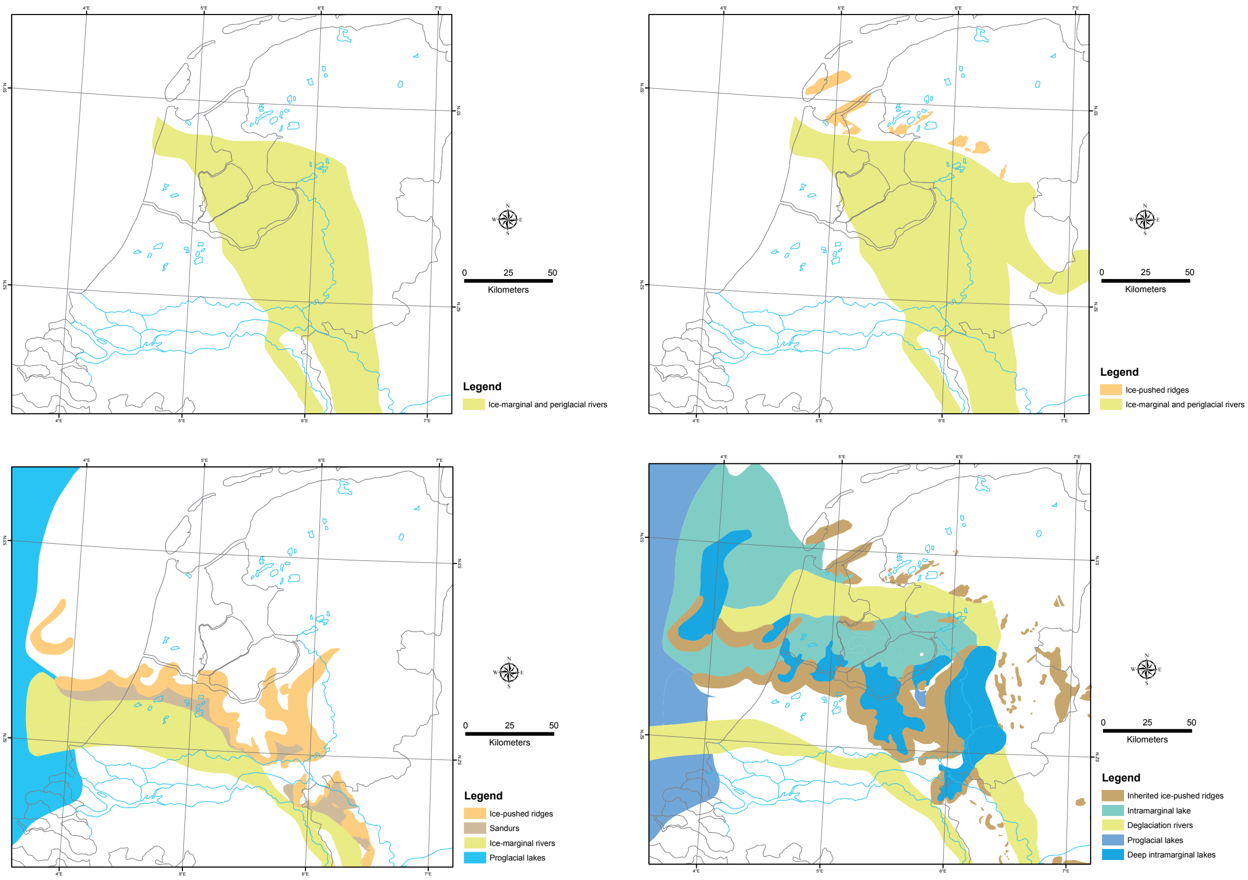


Figure 7.12
The phase model of Busschers from the GIS. As the the model of Busschers served as an important input for the GIS, this figures is quite comparable to figure 7.11

Phase 2: Phase two represents the maximum ice sheet extent. During this phase, 'unit S4' was deposited that yielded an age of 130-157 ka. The Rhine - Meuse system entered a proglacial lake located at the current North Sea, that reached heights similar to interglacial highstand.

Phase 3: During deglaciation several large deglaciation lakes formed, mainly in the deep glacial basins (e.g. 'Holland Lake'. The IJssel Valley was filled in with deposits from the Rhine 'unit S6' just before the onset of the Eemian. Deep incision occurred of the former proglacial Rhine - Meuse river plain by the Meuse during and after lake drainage ('Unit S5'). After lake level drop, smaller lakes remained in the Northern Netherlands (the glacial basins) until the onset of the Eemian transgression.

Discussion
This phase model integrates data from the proglacial and deglacial situation in the Dutch part of the study area, the glacial situation is considered as a boundary condition. Most of the elements of this model were also incorporated into the new phase model (compare figure 7.11 and figure7.12).
Phase 2 lumps many features (both ice-pushed and Hondsrug) it is too crude for a glacial phase model. Also, the formation of 'Unit S5' after 'Unit S4' and the evolution of the proglacil lake into Holland Lake is part of phase 3, whereas it started to form halfway/late within phase 2 (Busschers, Cohen, pers. comm.)

8. The updated phase model

The different classical phase models contain observational data and interpretations that provide valuable elements for the GIS database. Most of these phase models have a monothematical approach, or integrate only several aspects of the glaciation (figure 7.1). They are mainly based on an extrapolation of results and intepretations of extensive research in a local area (table 7.1). To unify the elements of the classical phase models and the local studies described a new framework is required, and therefore an updated phase model. This phase model considers the various glaciogenic geomorphological features in the study area. It uses the definitions for 'a phase' as used in chapter 2.1 and the GIS to store the phasing. It should be restated here that the transitions between these phases are transient, and not abrupt. The updated phase model combines elements and lines of reasoning from different existing studies, as described in chapters 6 and 7.

The updated phase model consists of six phases over the combined glaciation and deglaciation. During phase one the ice advances from the north up to the topographical obstacles in Germany. In phase two the ice front advanced from the northeast into the northern part of the Netherlands. Phase 3 corresponds to the maximal extent of the ice, phase 4 represents a dead ice field with a last ice stream marking the transition of glaciation to deglaciation. During phase 5 and 6 deglaciation takes place, two readvances occur in Germany corresponding to the 'Warthe Substage'.

8.1 Phase 1

During this phase, ice advanced from the north into the research area. This is confirmed by fabrics that are oriented from the NNW-NNE in eastern Lower Saxony (chapter 6.1 - Ehlers, 1990) and the NNW-SSE oriented glacial lineations of the Syke Geest. As the morphology of the Syke Geest is still preserved it is assumed that no significant ice activity occurred here after this phase. Probably, the relatively small 'Rehburg' ice-pushed ridges east of the Dammer Berge were formed as well. In Germany, the position of the northern rim of the Weserbergland coincides very well to the presence of the Rehburg line (Van der Wateren, 2003). Here, the required décollement layers reach a relatively shallow depth (figure 5.9). The ice subsequently overrode these ridges (cf. Meyer, 1987) and finally reached the high bedrock obstacles of the Weserbergland, where the ice stagnated. More to the west (in the Dutch area) the extension of this ice front becomes uncertain, because most of the evidence disappeared in later phases. It is assumed that the ice lobe diverted towards the ice margin, flowing towards the SW in the Netherlands.

During the onset of this phase the Weser, was still able to drain between the Wiehengebirge and the 'Maarleveld line'. At the end of this phase the Weser was probably blocked and deflected towards the west to follow the ice margin. A significant amount of meltwater was routed though sandurs and the ice-marginal Weser. The best preserved large sandur complex of this phase is the Hümmling in Germany, which probably formed between two ice lobes of the ice front (Schwan & Kasse, 1997). Underneath the till plateaus in Lower Saxony also thick meltwater deposits (10-30 m) are preserved (Vorschüttsande - chapter 6.4 & 6.5). Most likely, these deposits do not only represent sandur deposits, but also deposits from these ice-marginal rivers. In the northern Netherlands the downstream continuation of these deposits have not been described as such. In the northern Netherlands, gravelly top units within the Peelo Formation may represent its downstream continuation.

8.2 Phase 2

The flow direction of the ice changed during this phase, the ice front came mainly from the north-east instead of the north (figure 8.2). This change in ice flow direction is probably a combined effect of the shift of the ice divide at ice sheet scale towards the east, as happened during the Weichselian ice cover

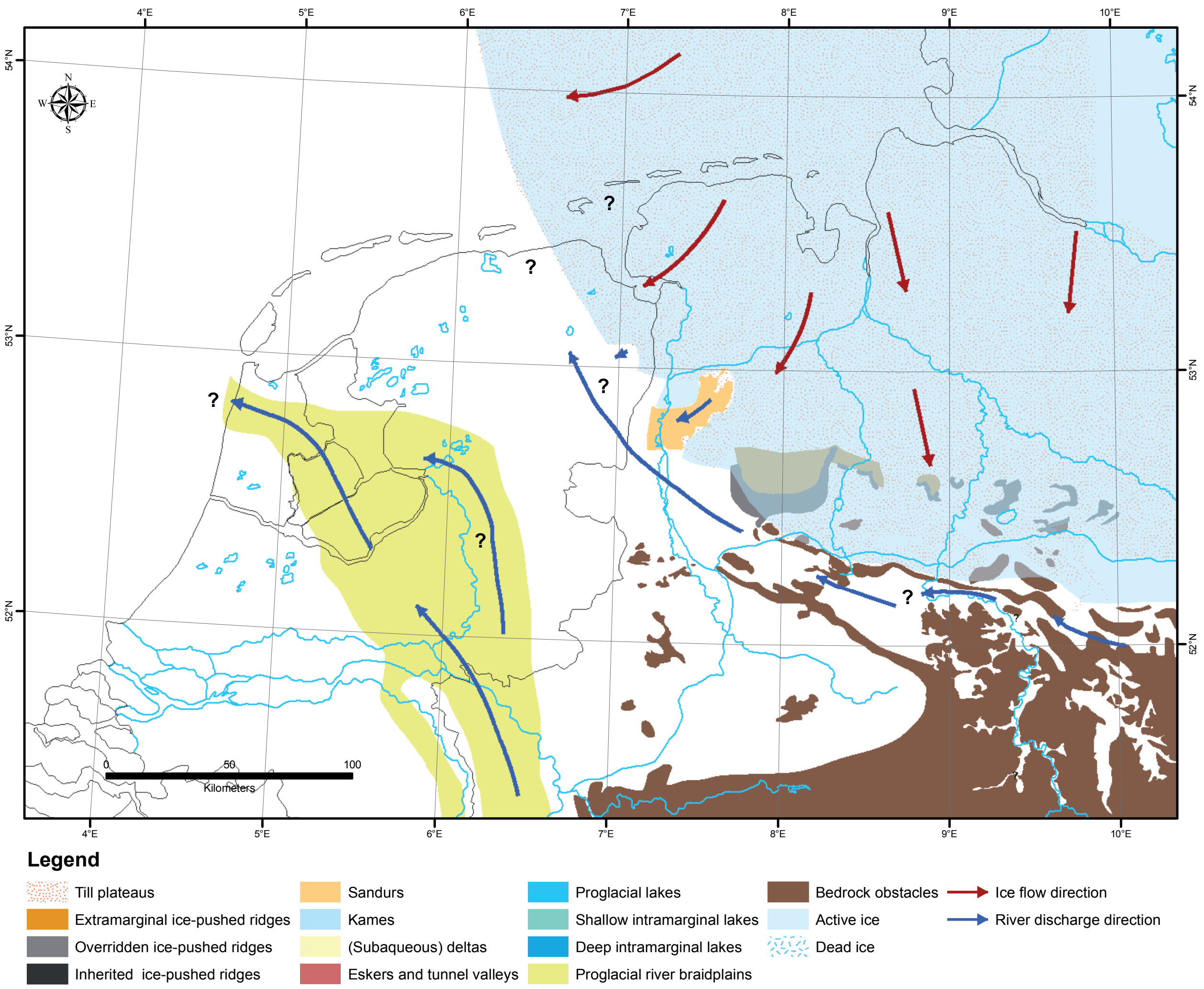


Figure 8.1
During the first phase the ice front enters the study area from the north

(Ehlers 1990ab; Houmark-Nielsen, 2010 - chapter 4.3.3) and the local effect of ice stagnation on the Weserbergland (see phase 1). The flow direction is evident from NE and NNE oriented fabrics and flutes in several till plateaus in northern Lower Saxony and Drenthe (chapter 6.1). Preserved truncated NNW-SSE lineations from phase 1 on the Syke Geest indicate that over that area the ice had become dead ice (inversion method explained in chapter 2.2 and 5.2.3).
At the ice front, the western 'Rehburg' ice-pushed ridges were formed (Dammer Berge, Twente ice-pushed ridges). In the initial proglacial situation, sandurs had formed here that are partially incorporated in the ice-pushed ridges and for the other part, overridden (Bruinehaar-Dinxperlo - Itterbeck sandur). The smaller ice-pushed ridges on the line Coevorden-Steenwijk-Gaasterland-Wieringen-Texel were formed as well (cf. Rappol, 1991). The interaction between the local substrate and the ice is considered the dominant factor for their formation. Consequently, although the ice-pushed ridges did form within one single phase, they did not necessarily form exactly simultaneous (chapter 5.3). The ice-pushed ridges in the west (Texel- Steenwijk) are younger that the ice-pushed ridges in the east (Dammer Berge), as the ice front had to pass a longer distance towards the western margin of this phase. The ice-pushed ridges east of the Dammer Berge were probably even already formed in phase 1. The former implies that the ice-pushed ridges in the east could already have been overridden while the ridges in the western part were just being formed.
Striking is the absence of the aeolian periglacial depostis of the Drachten Member east of the Hondsrug area in the northern part of the Netherlands, whereas they do occur west of this region (figure 3.12). This implies that the major part of the ice-marginal discharge connected to the Rhine

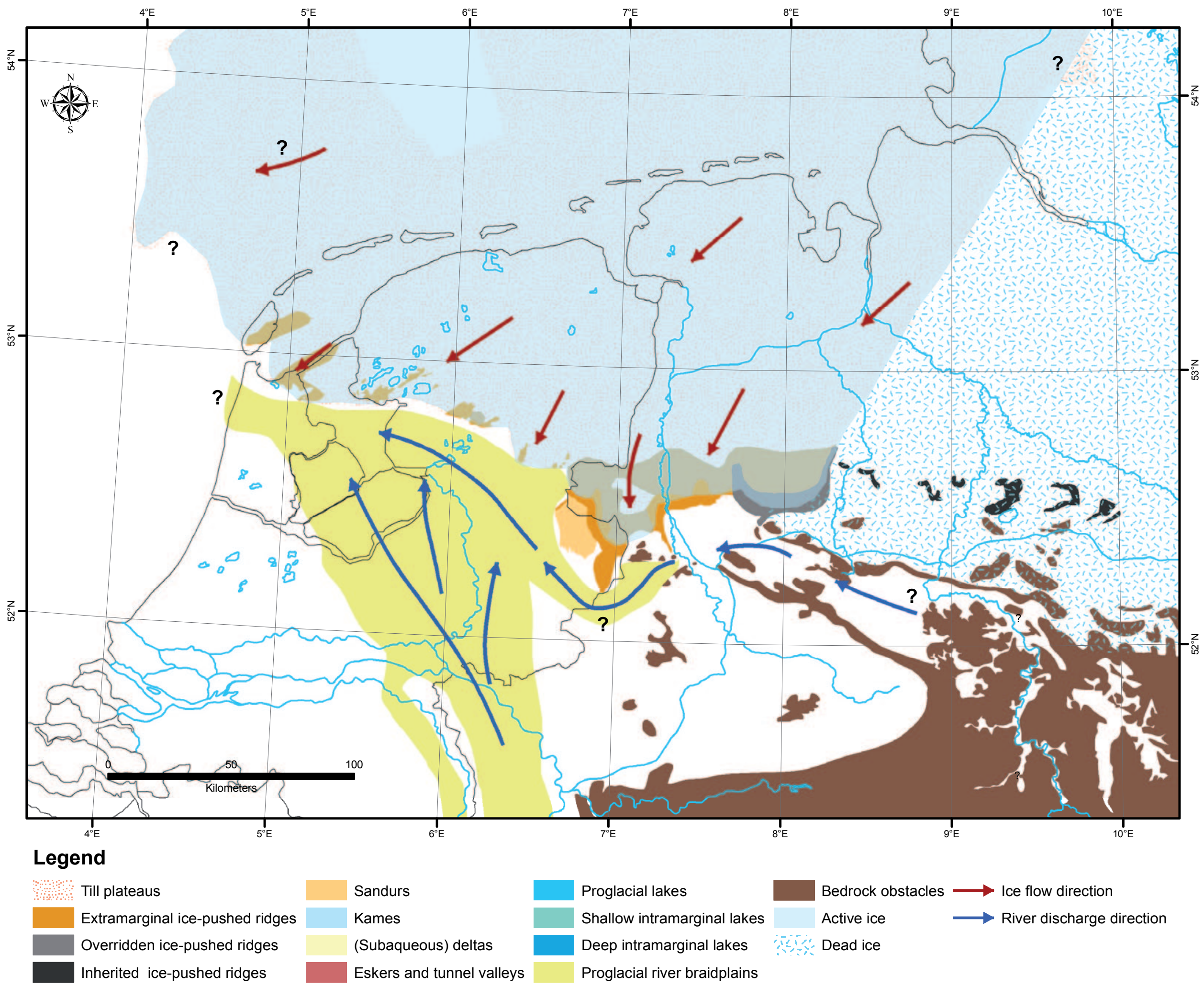


Figure 8.2
During the second phase a dead ice field forms in the eastern part of the study area and the ice progrades from the northeast.

when the ice front arrived in the Netherlands. This explains both the preservation of the aeolian deposits and the absence of ice-marginal river deposits in large parts of the northern Netherlands. In the central part of the Netherlands pradolinas routed towards the west entering the river Rhine that was present in the central part of the Netherlands (chapter 6.5, chapter 7.4.2 - Busschers et al., 2008). This aggrading river system filled up the incised river valleys yielding a flat topography. This implies a high base level, most likely induced by the rising lake levels of the proglacial lake North Sea. This river system caused permafrost to be nearly absent. Sediments from this ice-marginal river system have been OSL dated at 168 +/-19 ka (Busschers et al., 2008), which is assumed to coincide with the age of this phase. The presence, location and nature of the upstream ice-marginal river system of the Weser is uncertain. Up to now, no sediments have been recognized west of the Itterbeck sandur which contains some admixtures. This may be an indication for the initiation of the damming of the Weser and the formation of a proglacial lake in the Weserbergland from this stage onwards. It could also be that the ice-marginal sediments are present around the Dutch/German border near Twente, but not recognized as such. Therefore, these sediments underneath the Saalian glacial sediments have to be reinterpreted (Busschers, ongoing work).

8.3 Phase 3

During this final phase of ice advance, flow directions as in phase 2 advanced the ice front towards the south. The large ice lobe that was covering the north of the study are in phase 2 split in two distinct ice lobes with two main directions. One part headed mainly westward towards the proglacial lake North

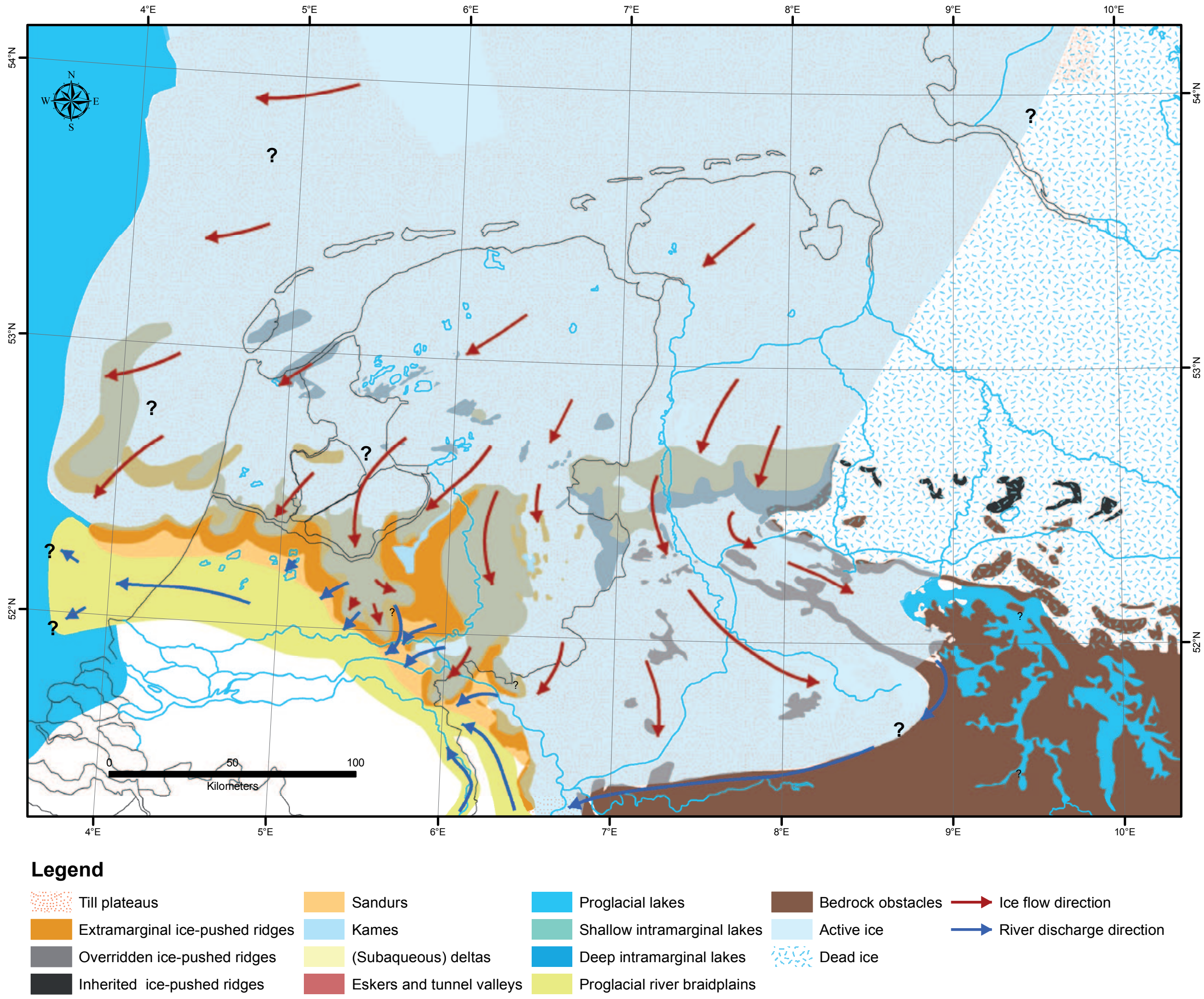


Figure 8.3
The maximal extension of the ice in the study area. Large proglacial lakes formed in the North Sea and the Weserbergland.

Sea. The other headed into the Münsterland Embayment (mainly southward). Between these main lobes, the resistance of the coarse Pleistocene deposits caused inhibition of the ice front progradation in this area, and the largest ice-pushed ridge complexes were formed.
The ice-pushed ridges that had been formed during the previous phases (Rehburg line) were overridden during this phase (cf. Van den Berg & Beets, 1987; Rappol, 1991 – see discussion chapter 7.1.3 and chapter 7.3.2). As in the previous phases, the German Weserbergland bedrock was an important obstacle for the prograding ice front. It caused the ice flow to deflect around its western tip, making it concentrate between Enschede and Osnabrück (cf. Skupin et al., 1993- chapter 7.2.1). The ice front prograded into the Münsterland Embayment where it diverted strongly. Locally, the ice was slightly deflected by the Baumberge and Beckumerberge (cf. Skupin et al., 1993). The ice advance stopped on the southern fringe of the Munsterland embayment, where topographical obstacles were present. In the western part of the study area hardly any sandurs and push moraines were formed. Here, the ice front entered the proglacial lake North Sea (chapter 6.6.2) causing sheet calving. This probably yielded a pulling force for the ice towards the west (general mechanism described in chapter 4.6).
Along the line Haarlem-Utrecht-Nijmegen-Düsseldorf major ice-pushed ridge complexes were formed, that mark the maximal extension of the ice in the Saalian (figure 8.3). In the central part of the Netherlands relatively high ice-pushed ridges and deep glacial basins were formed on the line Haarlem-Utrecht-Nijmegen-Düsseldorf (De Gans et al., 1987; Van den Berg & Beets, 1987). This morphology reflects a strongly lobed ice front, especially in the central part of the Netherlands. This

striking morphology was formed in a relatively thick and coarse unconsolidated substrate where an increased resistance caused a thickening of the ice lobe and the formation of large ice-pushed ridges (chapter 5.3.1 and 7.1.3 - Van den Berg & Beets, 1987). The glacial basins generally have a NE-SW orientation, in accordance to the flow main direction of the ice front. The orientation of the IJssel Valley and the Geldersche Vallei is somewhat peculiar. Here, two large ice tongues followed the Rhine and Meuse courses present in phase 1, which left a relatively coarse substrate. The ice-pushed ridges line up with the NW edge of the Peel Block (De Gans et al., 1987; Van Balen et al., 2005). This block acted as a minor topographical obstacle, but also as a hydrological discontinuity, which together have blocked further continuation of the weakened ice front. During and immediately after ice-pushed ridge formation, large meltwater valleys cut into them, that fed small sandurs in front of the ridges. A well-known example is the Darthuizerpoort in the Utrecht ridge. Probably major corridors formed in the Gooi region, were nowadays only small remnants of ice-pushed ridges are present. Also in the Veluwe some major meltwater corridors can be recognized, feeding the Schaarsbergen Sandur complex (compare figure 6.18 and figure 8.3). These large sandurs formed along the line of ice-pushed ridges from the North Sea to Düsseldorf, feeding the ice-marginal rivers with water and sediment. OSL-dates from the Rhine-Meuse ice margin sandur deposits yield 130-157 ka and ~150 ka (Busschers et al., 2008), which are believed to be realistic ages.
Locally, the sequence of the advancing of the ice lobes within this phase could be determined. For example, the Arnhem ice-pushed ridge is younger than the Eastern Veluwe ice-pushed ridge (Bakker, 2006 - chapter 7.1.1). From the truncation of the Zuidwolde ridge (chapter 6.2.3) and the structures of the Woldberg (northern tip Veluwe) may be judged that the ice front progradation towards the North Sea proglacial lake was somewhat longer active that the ice flow towards the south. These differences are considered to be too local to correlate to a regional overview, and are therefore not included into the phase model (see chapter 2).
During the progradation of the ice front towards its maximal extension the westward flowing Weser was dammed off by the ice, which caused the formation of the Weser Lake (chapter 6.6.1 - Winsemann et al., 2009; 2010). On the northern side of the Veluwe kames were formed. The Rhine was truncated and discharge joined a newly formed ice marginal river plain, running WNW through the central part of the Netherlands. The large ice-marginal river built out a delta ('Unit S4') into the proglacial lake North Sea that filled the non-glaciated part of the North Sea basin (chapter 6.6 - Busschers et al., 2008). From the amount of sediment deposited in this delta can be deduced that it was active for a relatively short time ~2 ka, and from its geometry the lake level was estimated to have been around interglacial sea level (Busschers et al., 2008). Also, the narrow Rhine corridor in the LRE implies a non equilibrium situation, i.e a short existing time.

8.4 Phase 4

During this phase the largest part of the ice field stagnated, only one single ice stream from the NNW occurred in the Hondsrug area. This ice stream will be referred to as the Hondsrug Ice Stream (HIS). Phase 4 partially reproduces 'phase 3' the HIS of Rappol (1987); and phase 2 by Van den Berg & Beets, 1987 - chapter 7.1.3, 7.3.2. The HIS markes the beginning of deglaciation. It affected the western edge of the Hümmling (cf. Schröder, 1978; Rappol, 1991; Speetzen & Zandstra, 2009) and continued in eastern Twente (cf. Kluiving et al., 1991) into the Münsterland Embayment (cf. Van den Berg & Beets, 1987; Skupin et al., 1993). This relatively fast ice stream caused large volumes of ice to be transported from the region north of the Netherlands in the North Sea region (mechanism described in chapter 4.6). Because the ice stream routed ice away from this area, it caused a relatively rapid initiation of deglaciation in its source area (Plasschier et al., 2010).
The direction of the HIS is deduced from the fluted morphology of the Hondsrug (chapter 6.1 - Van den Berg & Beets, 1987) and in fabric analyses (Rappol, 1991). Tills delivered by the HIS are distinct from other phases because of the characteristic eastern Baltic (Area 1, table 6.2) erratic assemblage.

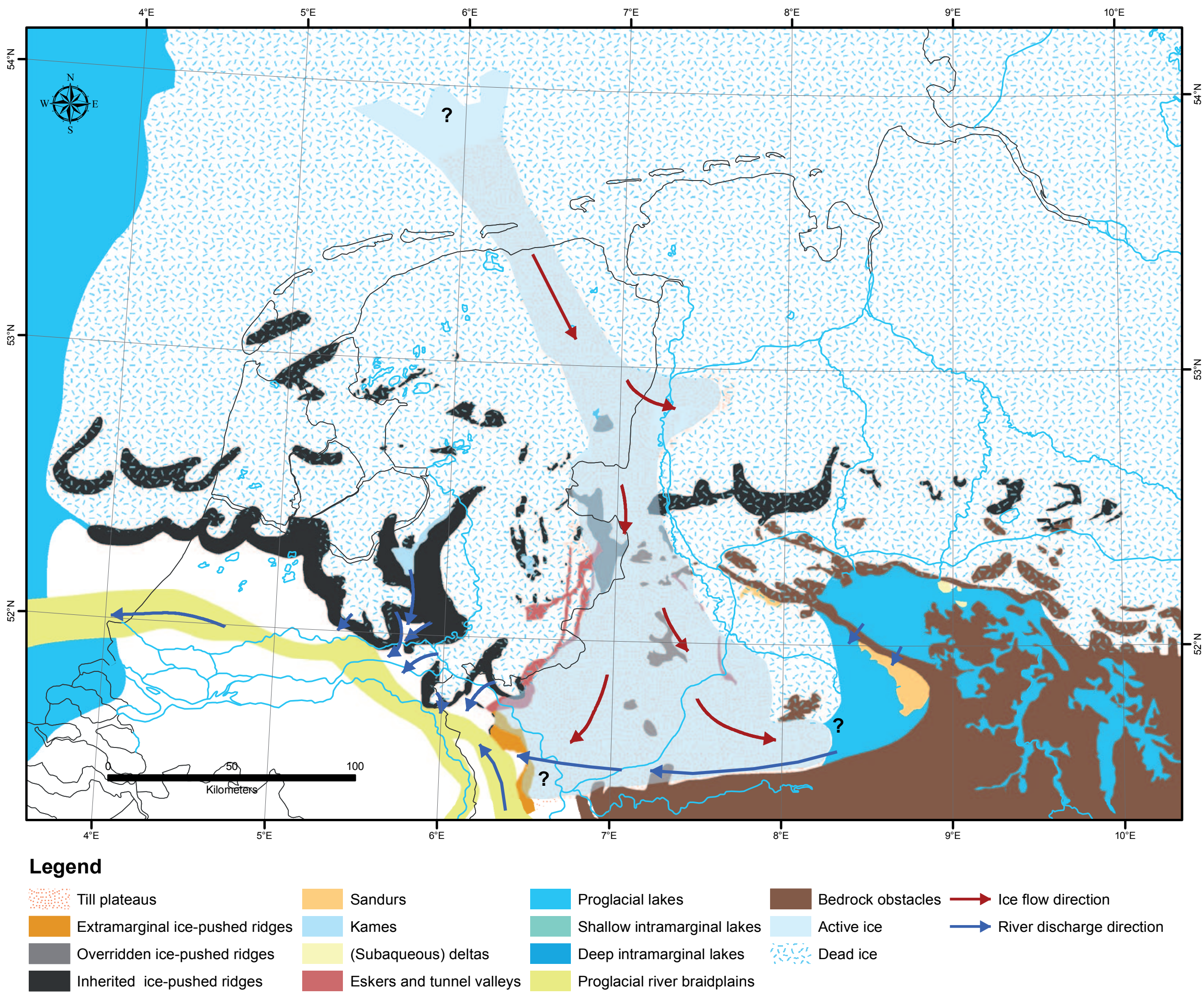


Figure 8.4
A final ice stream marks the transition from glaciation to deglaciation.

This was used to trace this ice stream into the Munsterland Embayment. As the morphology from earlier phases was clearly preserved dead ice fields must have been present in the adjacent regions (Rappol, 1987). In the Münsterland Embayment the contact between the dead ice and the active ice of this phase was marked by the Achterhoek-Twente tunnel valley system and the Münster Kieszandzug (Skupin et al., 1993). Probably, no major changes occurred in the ice-marginal river patterns relative to the previous phase. In the Münsterland Embayment, the ice retreated and overspilling from the Lake Weser occurred into the newly formed Lake Münsterland. Spillage from this lake was probably routed partly over, or partly underneath the ice sheet, towards the LRE. It explains the severely eroded the ice-pushed ridges, the so called 'Inselberge' in the very south of the HUND-line (chapter 6.2.8).
A major issue of debate is how the Hondsrug ice stream was induced. According to Rappol (1987; 1991) the NNW direction of the HIS implies that it was induced by the collision of the Fennoscandian ice sheet and the British ice sheet, from this phase onward. Busschers et al. (2008) conclude that the British and the Fennoscandian ice sheets must have been in contact, because they hold up the lake in the southern North Sea, which already occurred in an earlier phase. The proglacial lake in the North Sea is not evidence that the collision of the Brithish and the Fennoscandian ice sheet. In fact ice cap collision as a trigger of the HIS is very unlikely because such a collision would yield a much larger ice stream than the HIS. Passchier et al. (2010) state that drainage of the proglacial lake North Sea and change in subglacial bed conditions in the North Sea region induced the HIS. This view is not followed here, as there is no convincing evidence for these mechanisms. Besides, only the upstream part of the ice stream is considered instead of all the involved areas. An alternative explanation grew from

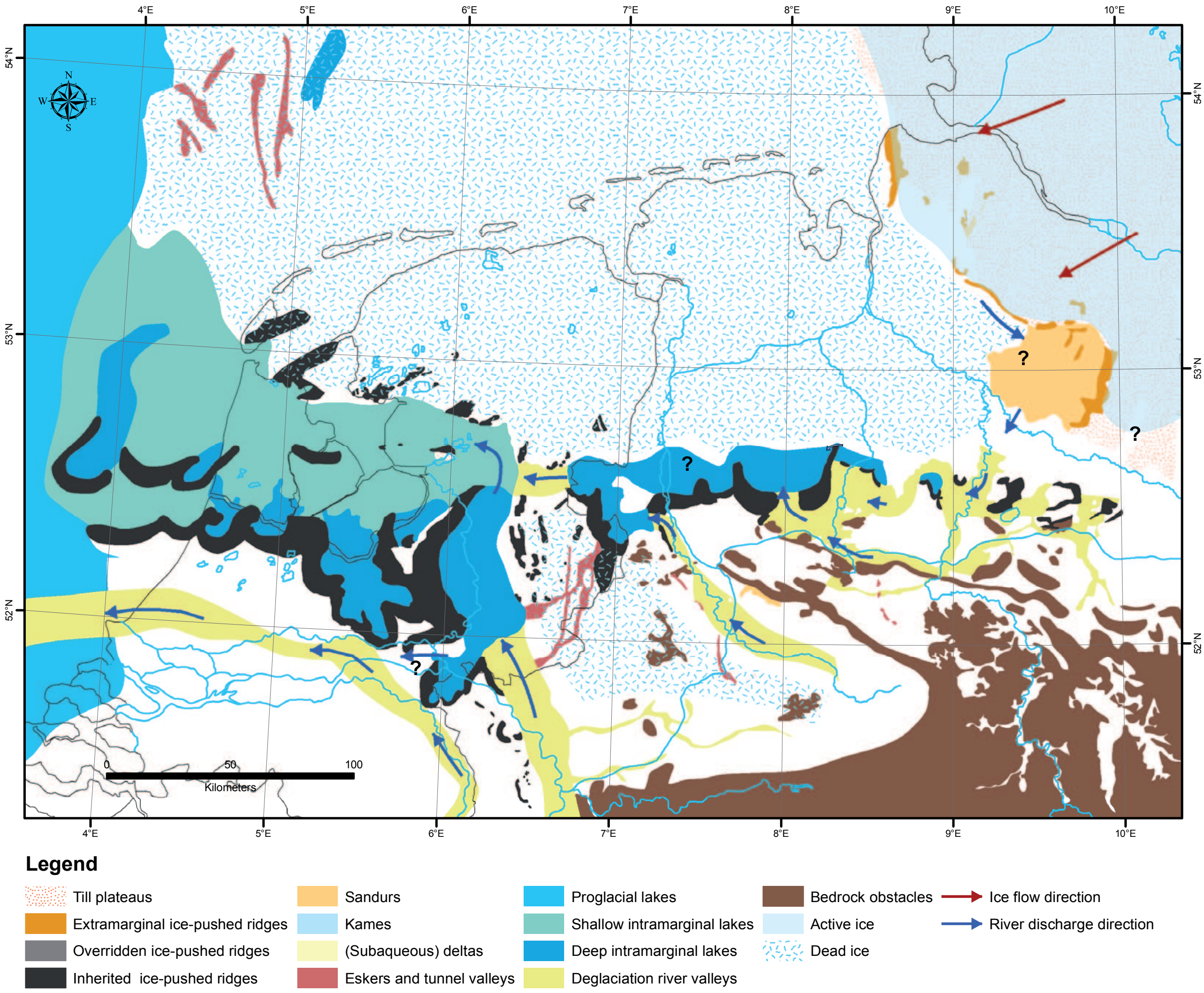


Figure 8.5
The first phase of deglaciation: intramarginal lakes and deglacial river valleys form mainly in the southern region of the study area. In the Elbe region a readvance of the ice front occurs.

the regional overview obtained in this study. We propose the HIS to mark ice stream activity within a stagnant ice field, at the turning point of glaciation to deglaciation. The ice lobe in this region can be considered as a dome. When the ice becomes stagnant, ice can seek its way trough the weakest area. The centre of the dome, immediately north of the Wadden Sea, a relatively large amount of ice was present. Under its own gravity and ice flow, it was able to intrude the weakened dead ice to the south along corridors of favourable subsoil. Probably, wet proglacial conditions in the south eastern part of the Münsterland Embayment (Winsemann, Meinsen, in prep) caused major ice sheet calving and lifting of the ice lobe from its bed, acting as a pulling force. The latter mechanism is a well known phenomena at ice sheet/ lake interactions (chapter 4.6). It may have acted in the NW of the study area too, allowing the proglacial lake North Sea to expand in former glaciated areas.
The eastern Baltic erratic assemblage in an ice stream from the NNW is somewhat paradoxical. It may be explained by the change in shift of the ice divide (chapter 4.3.3) in Scandinavia at the onset of this phase, causing the source area of the erratics to shift towards the east or they can be the product of reworked tills that were present in the North Sea region since the Elsterian glaciation.

8.5 Phase 5

During this phase, large amounts of ice melted in the southern part of the research area (figure 8.5.). The presence of the proglacial lake indicates that ice was still present in the North Sea NW of this study area. Therefore, it is assumed that during this initial deglaciation phase the meltwater was routed

to the west and south-west.
The Rhine diverted into the IJssel basin where a lot of sediment could be deposited. The meltwater from this lake drained into the large Holland Lake in the Western Netherlands (chapter 6.7.2), which was directly connected to the proglacial lake on the North Sea. North of the Weserbergland a series of meltwater channels routed towards the west trough the Vecht and Hunze valley.
The large amounts of meltwater routed trough wide deglaciation valleys which eroded substantial parts of the glacially formed features. At critical positions in the drainage network these rivers were interrupted by deep deglaciated tongue basin lakes (chapter 6.7). North of the Weserbergland a series of meltwater channels routed trough the Quackenbrück glacial basin towards the Nordhorn glacial basin. In the Münsterland Embayment probably a major channel was present along the Teutoburgerwald to the northwest that also entered the Nordhorn Basin. The Nordhorn Basin spilled westwards into the Vecht valley. More to the south, the Rhine entered the deep IJssel Basin (Van der Meene, 1977; Busschers et al., 2008), which also spilled into the Vecht valley. In the southern part of the IJssel Valley, Rhine delta deposits were formed (Kreftenheye Formation, 'Unit S6'), in more distal positions prodeltic muds formed. In the other glacial basins lacustrine deposits were formed (chapter 6.7.3). The Vecht drained into the Holland Lake (Beets & Beets, 2003; Busschers et al., 2008), which was directly connected to the proglacial lake North Sea. The water levels in these lakes are estimated at -20/-30 relative to sea level, after a significant drop due to erosion in the the Dover Strait and Southern Bight.
In northeastern Lower Saxony a reactivation of the ice front occurred between the Elbe and Weser. This was evident from the till stratigraphy (till layer was separated from the Drente till by meltwater deposits – chapter 6.1.6) and the occurrence of readvance ice-pushed ridges with older tills incorporated (chapter 5.3.3 and 6.2.6). The Lamstedt and Altenwald ice-pushed ridges run parallel to a series of N-S trending salt domes (Kuster & Meyer, 1979), which pushed up layers that could act as a suitable décollement (Van der Wateren, 2003). Some of these ice-pushed ridges were overridden and large sandurs formed on the Lüneburger Heide (chapter 6.4.3). Fabric analysis indicated ice flow from the northeast. This readvance can be linked to the Middle Drenthe phase of Ehlers (1990a;b – chapter 7.3.1). Some of these ice-pushed ridges were overridden and large sandurs formed on the Lüneburger Heide.
The exact timing of both the readvance phases and the deglaciation is not certain. It could very well be that the development of the deglaciatial situation during phase 5 and 6 took much longer than the two readvances of the ice front. However, the relative order of sequences (first deglaciation in the south, than in the north and the order of the two readvances) is supported by convincing evidence.

8.6 Phase 6

During this phase the deglaciation continued. The ice contact between the Scandinavian and the British ice sheet disappeared, causing the final drainage of the proglacial lake North Sea. This also caused the Holland Lake to be drained and to disappear, approximately 1 ka after the start of its infilling (Beets & Beets, 2003). Due to the ice-free conditions in the north the meltwater could also be discharged toward the north trough the Hunze valley and the Weser-Aller valley (Toucanne et al., 2009). The Hunze valley formed after overflow of a glacial basin from the Rehburg line (Van den Berg & Beets, 1987). In Germany, an eastern branch of the Hunze is present that can be traced between the Oldenburg and Ostfriesland (Speetzen & Zandstra, 2009). The Weser-Aller valley can be traced in the present landscape as a large elongated depression of several tens of kilometres wide (figure 6.6). It was formed during the deglaciation of the Drenthe stage when it drained a substantial part of the eastern part of the research area and probably the southwestern edge of the whole ice sheet. It was enlarged when it became a ice-marginal river during the Warthe readvance stage (Meyer, 1983; Ehlers et al., 2004). This drop of the erosion base and the low sediment content of deglaciation rivers draining the glacial basins (e.g. Vecht and Hunze) caused deep incision (Van den Berg & Beets, 1987).
Somewhat further to the east, around Hamburg (figure 8.6), a small readvance of the ice front occurred

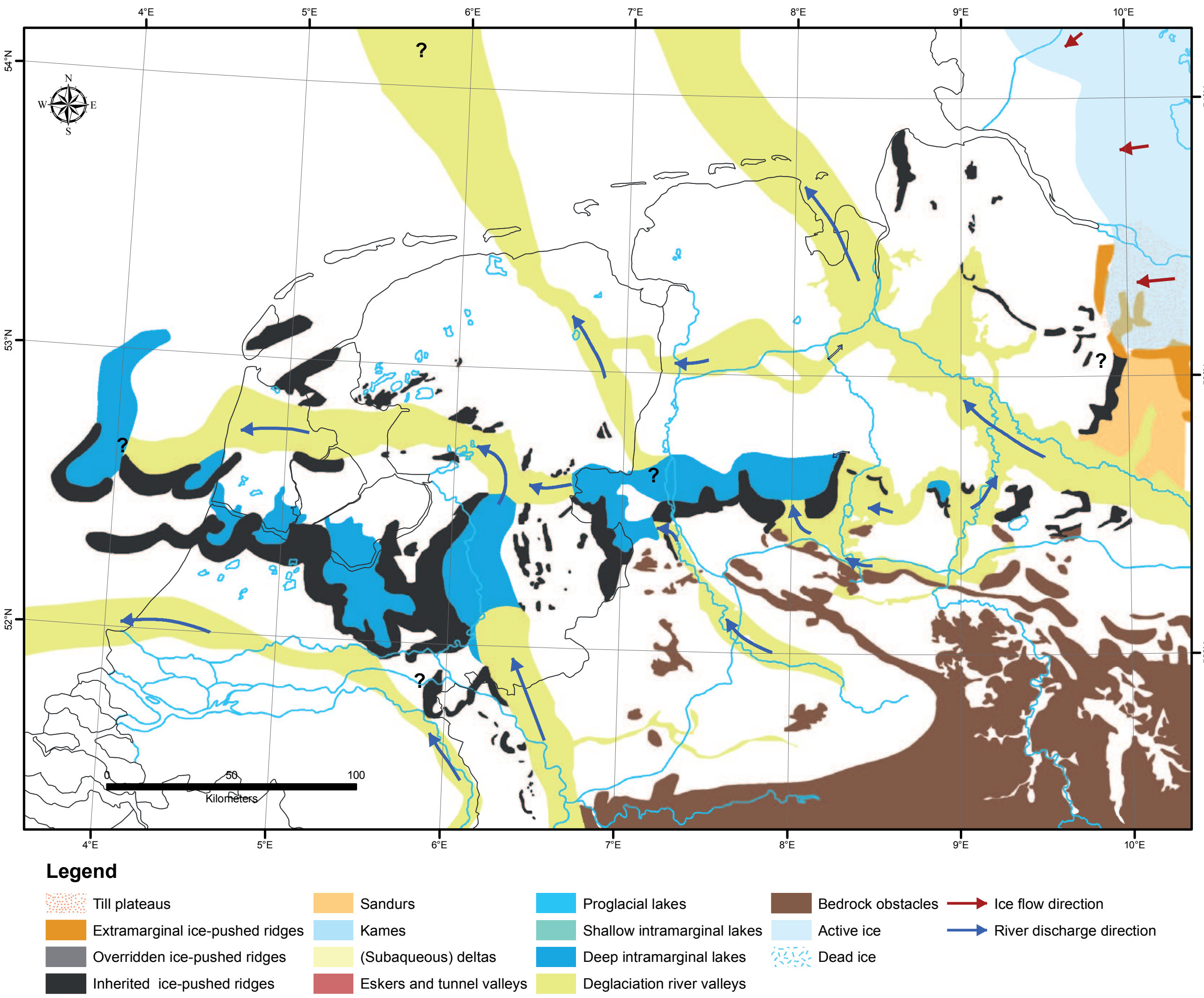


Figure 8.6
Full deglaciation: the meltwater also drains towards the north. A readvance of the ice front occurs in the Elbe region, which can be linked to the Warthe stage.

as described in chapter 7.3.1. This readvance can be linked to the Younger Drenthe phase or Warthe (Ehlers 1990a;b; Meyer, 1987; 2005). The ice front prograded from an eastern direction, probably related to the further shift of the ice divide towards the end of the glaciation (analogous to the ice flow at the end of the Weichselian glaciation - chapter 4.3.3). There is no evidence for interglacial/ interstadial conditions between these two ice advances (Ehlers et al., 2004). After a few ka, the Eemian started in this region, no major hiatuses are evident between the glacial deposits and overlying Eemian (Van Leeuwen et al., 2000; Busschers et al., 2008).

8.7 Concluding remarks: new insights from the phase model

The updated phase model provides a regional overview that yields some new insights and some starting points for new research:

- The presence of topographical obstacles and conditions of the substrate (yielding resistance for the ice flow) are the most important factors for ice flow and ice stream patterns.
- Long lines of ice-pushed ridges were formed diachronously. During the onset of glaciation the ice front reached the eastern part of the Rehburg line earlier than the western part
- The striking absence of meltwater deposits underneath the tills in the central north of the Netherlands

indicates either merging of the ice-marginal Weser with the Rhine or trapping of Weser in lakes upstream.

- Apparent absence of meltwater deposits around the Twente region could mark the initiation of the damming of the Lake Weser, however renewed study of sediments underneath the tills is required to validate if the deposits are really absent or were misintepreted.
- While glaciation of the Netherlands and the Münsterland Embayment was taking place, a large dead ice field must have formed north of the Weserbergland.
- Just before the onset of the maximal extension two large lobes formed towards the proglacial lake North Sea and the Munsterland embayment, where wet proglacial conditions occurred and large topographical obstacles lacked. Most likely, substrate conditions (thickness and coarseness of substrate, presence of geothermal sources) enhanced relative fast ice flow. Substrate conditions in the central part of the Netherlands (e.g. presence coarse deposits) caused a resistance for the ice front.
- During the maximal extent the ice front was marked by: the proglacial lake North Sea (where ice berg calving occurred), the ice-pushed ridge-sandur complexes on the line Haarlem-Düsseldorf, topographical obstacles (southern rim of Münsterland Embayment) and the Lake Weser.
- The dimensions of the Rhine deposits 'Unit S4' formed during the maximal stage suggest a relatively short duration (~2 ka) of the maximum stage (phase 3), at least before erosive lowering of the Dover Strait and South Bight and well before the formation of the Holland Lake stage (stage 5).
- Evidence from the Münsterland Embayment suggests the presence of a lake in the Münsterland draining towards the west into the Rhine. The exact pathway of these large amounts of meltwater, the connection to the Dutch situation and implications for the chronology requires further research.
- The Hondsrug ice stream marks the transition between glaciation and deglaciation. It was most likely the consequence of a mass surplus in the northern part of the study area, the presence of geothermal sources in Drenthe and the pulling capacity of the Münsterland Lake.
- During and after the formation of ice-pushed ridges large meltwater erosion corridors were formed in them. The best example are the severely eroded ice-pushed ridges in the LRE, their erosion may be linked to the drainage of the Lake Münsterland.
- During deglacation, large glacio-fluvial valleys formed that discharged huge amounts of water released by the retreating ice sheet margin. At critical positions, the drainage network was interrupted by deglaciated tongue basin lakes.
- The timing of the events in the deglaciation can be improved by using warve chronology in the infilling of the glacial basins and OSL dating of sandur deposits formed during the reactivation of the ice front.

Conclusions

In this research the sequence of glacial events in the Drenthe substage in the Late Saalian (MIS 6, around 170,000-150,000 years ago) was newly reconstructed for NW Germany and the Netherlands. The newly constructed phase model recognises three phases towards maximum ice-sheet extent, one transition phase and two deglaciation phases. From this reconstruction several conclusions could be drawn:

- In order to reconstruct the Saalian glaciation and deglaciation, datasets considering different aspects of the glacial, proglacial and deglacial situation have to be integrated. Besides, the existing (mainly local) studies have to be integrated into a regional framework. For this, a strict separation of the observational data and their interpretation was required. In this way, the matching and conflicted points could be compared, the best interpretations were incorporated into the phase model.
- Implementation of glacial, proglacial and deglacial phenomena in a GIS provided a solid dataset for observational data, interpreted classical phase models and updated phase model reconstruction. Due to its structure the GIS provides a data base in which the features are equally weighted.
- The regional scale ('zoom out') of the reconstruction yielded new insights of the processes that occurred during the glaciation. The main insights, a new constructed iterative phase model, and some recommendations are listed in chapter 8.
- The GIS-dataset and integrated, updated, new phase model provides a valuable framework for future research. It could focus on mechanisms of interaction of glacial dynamics with substrate processes (waterflow, heatflow) and conditions (lithology, thickness). The model also provides an glacial context for ongoing or future proglacial studies.
- The regional scale of the reconstruction allows inter-comparing feature sets that have a comparable substrate setting. The regional scope can be extended along the ice margin towards regions that now served as a boundary condition (Dover area, the situation towards Eastern Germany)
- The dataset and phase model also serve to focus dating sampling strategies, and vice versa accumulated dating results ('swarms of dates') may help to validate the model. Another way to validate the model is the analysis of clay minerals.

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