Modelled ice-sheet margins of three Greenland ice-sheet models compared with a geological record from ice-marginal deposits in central West Greenland

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ABSTRACT. Ice-sheet modelling is an essential tool for estimating the effect of climate change on the Greenland ice sheet. The large spatial and long-term temporal scales of the ice-sheet model limits the amount of data which can be used to test model results. The geological record is useful because it provides test material on the time-scales typical for the memory of ice sheets (millenia). This paper compares modelled ice-margin positions with a geological scenario of ice-margin positions since the Last Glacial Maximum to the present in West Greenland. Morphological evidence of ice-margin positions is provided by moraines. Moraine systems are dated by 14C-dated marine shells and terrestrial peat. Three Greenland ice-sheet models are compared. There are distinct differences in modelled ice-margin positions between the models and between model results and the geological record. Disagreement between models and the geological record in the near-coastal area is explained by the inadequate treatment of marginal processes in a tide-water environment. A smaller than present ice sheet around the warm period in the Holocene (Holocene climatic optimum) only occurs if such a period appears in the forcing (ice-core record) or used temporal resolution. Smoothing of the GRIP record with a 2000 year average eliminates the climatic signal related to the Holocene climatic optimum. This underlines the importance of short-term and medium-term variations (decades, centuries) in climatic variables in determining ice-margin positions in the past but also in the future.

INTRODUCTION

Ice-sheet modelling is an essential tool in evaluating the future response of ice sheets to climate change. Models need as many tests as possible to enable a judgement to be made on the performance of models. Every time a model passes a test, more confidence may be placed in model results for periods without control options such as the future. They should especially be evaluated for circumstances and time-scales in which they will be used (Oreskes and others, 1994).

The large spatial and long-term temporal scale of ice-sheet models limits the amount of data which can be used to test model results. Modelled ice-sheet geometry (span, topography, volume) can be compared with present-day geometry for the Greenland and Antarctic ice sheets. Modelled ice temperature can be tested against measured temperatures at deep drilling sites in Greenland and Antarctica.

The potential of the geological record to provide test data for ice-sheet models is often mentioned (e.g. Paterson, 1981; Hindmarsh, 1993). However, systematic comparisons between geology and ice-sheet models are not widespread. Data in geology are diverse in nature (morphological, sedimentological) and occur on a wide range of scales, in size and time. Not all types of geological data can be used in comparisons with ice-sheet model results. Based on the approach of Tatenhove and others (in press), only those geological features are used which relate directly to model output. The geological record used here is a sequence of dated moraines. Moraines are the morphological expression of the ice-sheet margin and therefore relate directly to modelled ice-sheet geometry.

Using a geological record for testing ice-sheet models
is useful because it provides test material on the time-scales typical for the memory of ice sheets (millennia). It also enables the analysis of transient behaviour of ice-sheet models.

The aims of this paper are:

1. To compare modelled positions of the ice margin of three Greenland ice-sheet models with geological evidence.
2. To evaluate the performance of the three models with respect to the geological record.
3. To discuss the differences in performance between the three models in terms of model characteristics.

GEOLOGICAL RECORD

For this paper we used the morphological evidence of ice-marginal positions as provided by moraine systems. Moraine systems are clusters of moraines which are traceable over large distances (10–100 km). Regional significance is attached to these systems because they are traceable over large areas and continue without being interrupted by topographic features such as valleys (e.g., Ten Brink, 1975). The positions of moraine systems (Fig. 1) are taken from various sources (for a review see Tatenhove, 1995; Tatenhove and others, in press). A detailed discussion on the methodology used in construct-

**Table 1. Deglaciation chronology of West Greenland. Division in groups is based on the availability and type of age determination.**

<table>
<thead>
<tr>
<th>Group</th>
<th>Moraine formation stage</th>
<th>Age in $^{14}$C year BP (± absolute range)</th>
<th>Age in calendar year BP (± absolute range)</th>
<th>Absolute range (calendar $^{14}$C year BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Hellefisk</td>
<td>16,000±3000</td>
<td>19,300–12,700</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sisimiut</td>
<td>12,300±1500</td>
<td>13,800–10,800</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Taserqat</td>
<td>9900±600</td>
<td>10,500–9300</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>Sarfartoq/Advedtloeq</td>
<td>8800±300</td>
<td>9200±600</td>
<td>9800–8600</td>
</tr>
<tr>
<td></td>
<td>Fjord</td>
<td>8300±300</td>
<td>8500±600</td>
<td>9100–7900</td>
</tr>
<tr>
<td></td>
<td>Umiivit/Keglen</td>
<td>7000±500</td>
<td>7300±600</td>
<td>7900–6700</td>
</tr>
<tr>
<td>C</td>
<td>Orkendalen</td>
<td>5900±300</td>
<td>6800±300</td>
<td>7100–6500</td>
</tr>
<tr>
<td>D</td>
<td>Minimum position</td>
<td>4000±900 (?)</td>
<td>4000±900*</td>
<td>4900–3100</td>
</tr>
<tr>
<td></td>
<td>Hypsithermal</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>End advance</td>
<td>Younger than AD 1750±100</td>
<td>260–110</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Neoglacial</td>
<td></td>
<td>200±30</td>
<td></td>
</tr>
</tbody>
</table>

*Geological model is developed assuming that the Hypsithermal minimum position of the ice sheet is 50 km east of the Orkendalen moraine system.
ing the geological scenario is given in Tatenhove and others (in press a).

Moraine systems were dated by \(^{14}\text{C}\)-dated marine shells and terrestrial peat, and were divided into four groups depending on the way ages were determined (Table 1). Groups A and D contain moraine systems without \(^{14}\text{C}\) dates. The relationship between relative sea level, which holds geological remnants with \(^{14}\text{C}\)-datable material, i.e. marine shells and moraine systems, is used to provide ages for group B. Group C is dated by terrestrial organic material found between morainic ridges. Although some \(^{14}\text{C}\) dates can be used to constrain the age of group D, this group does not give direct evidence of position, because it is presently covered by the ice sheet.

The ages of moraine systems are based mainly on the work of Ten Brink and Weidick for group B (Weidick, 1972; Ten Brink and Weidick, 1974; Ten Brink, 1975). The ages of the Umiit/Keglen, Orkendalen systems and group D have been discussed by Tatenhove (1995) and Tatenhove and others (in press b). The age ranges of the moraine system include the time period encompassed by the moraine system, the error related to the \(^{14}\text{C}\) dating itself and the uncertainty due to the inconclusive relationship between \(^{14}\text{C}\) date and moraine system. For the Hellefisk and Sisimiut moraines, no \(^{14}\text{C}\) dates are available within the research area and the assigned dates are based on arguments given in Tatenhove (1995). Moraine systems with a limiting \(^{14}\text{C}\) date close to the former ice margin (Fjord, Umiit/Keglen) have a relatively small age range.

Moraine-system age and associated accuracy ranges are interpolated to a 1 km\(\times\)1 km grid over an area of about 57 500 km\(^2\). Re-sampling to the area covered by a model grid point enables a quantitative comparison with modelled ice-margin positions.

**GLACIOLOGICAL MODELS**

The three models used in this study all cover the entirety of Greenland. An overview of the fundamental mathematical equations and model characteristics can be found in Fabre and others (1995) for the model denoted with “Fabre”, in Greve (1995) for the model “Greve” and in Huybrechts (1994) for the model indicated by “Huybrechts”. These models are based on the model discussed in Huybrechts and others (1991). The three models calculate the three-dimensional temperature and velocity.
field in the coupled mode to account for the temperature-dependent deformation of ice. The softness parameter of ice, which determines the rate of deformation, is a function of both temperature (all models) and the age of ice ("Huybrechts", "Greve"). Mass flow is by internal deformation and by basal sliding in the case of temperate basal temperatures (all models).

The models use the same data set for present-day bedrock topography (Lêtêguy and others, 1991). The models calculate isostatic bed adjustment with a time lag due to asthenospheric viscosity. The temperature evolution in the lithosphere is calculated up to a depth of several kilometres.

Present-day surface temperatures and accumulation rates are derived from Ohmura (1987) and Ohmura and Reeh (1991). The parameterization of surface temperature is in all three models treated as deviation from present-day values, using the ice-core record of GRIP (Fabre, Greve) or a compilation of the Dye 3 and Camp Century records for the period since the Last Glacial Maximum, after correction for elevation changes and origin of the ice (Huybrechts; Table 2) to estimate past values. No account is given for the spatial variability in time. The parameterization of accumulation is also expressed as deviation of present-day values. The Fabre model includes an orographic component, thereby creating some spatial variability in time in the accumulation pattern.

Sea-level change is ignored by the Greve model and is dealt with in a conceptual manner by the Fabre model (Table 2). The Huybrechts model uses the New Guinea record (Chappel and Shackleton, 1986) to force sea level. The grid size of the models is 20 km ("Fabre", "Huybrechts") and 40 km ("Greve"). The Greve model has used two different forcing functions. "Greve1" used a smoothed GRIP record (2000 year averages), while "Greve2" used the original, unsmoothed GRIP data. The position of grid cells of the three models is given in Figure 2.

RESULTS

The modelled ice margins of the Greve and Huybrechts models are generally older than the margin age derived

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**Fig. 2. Position of grid cells of three Greenland ice-sheet models (Fabre, Greve and Huybrechts).** The area covered is equal to Figure 1. Black cells denote the modelled position of the present-day margin.

**Fig. 3. Distance (in UTM) from west to east versus geological and modelled ages of ice-margin position (expressed in 10^3 calendar years).** The figure displays modelled ages for the x coordinates of the grid cells shown in Figure 2. It is assumed that the minimum extent of the ice sheet was 50 km behind its present position. The continuous line gives the average geological age of moraines.
from the geological record (Fig. 3). The average differences between the models and geology are 900 and 650 years, respectively. The Fabre model produces similar differences on the present-day offshore area, but it is in close agreement with the geological record onwards (Fig. 3). For this model, the average difference with the geological record is 400 years for the present-day land area. The deglaciation rate is reasonably reproduced by all models (Fig. 3).

As illustrated in Figures 3 and 4, modelled positions of the ice margin become closer to the geological scenario after about 10 ka BP or from 445 km UTM (Universal Transverse Mercator) eastward. Before this period, the difference between geology and model is much larger (Fig. 4). The modelled deglaciation rate is similar to the geological deglaciation rate for the land area between 445 and 560 km UTM or for the time period 10–6 ka BP (Fig. 3).

There are larger differences between the three models regarding the timing in reaching the present-day coastline. Moraines near the coast have an estimated age of 11–10 ka BP. The margin is at the present-day coastline at 18 ka BP in the Greve model (where it stays till 12 ka BP), between 14.5 and 12.5 ka BP in the Huybrechts model and between 12 and 10 ka BP in the Fabre model.

The Orkendalen moraines are close to the present-day ice margin and have an age of 7.1–6.5 ka BP (Tatenhove and others, in press a). Grid cells at the positions of these moraines have a marginal position at 7.5 ka BP ("Huybrechts") 8.5 ka BP ("Greve") and 6.4 ka BP ("Fabre"). The ice sheet retreated after the Orkendalen period behind its present position. In the Huybrechts model, this retreat comprises two grid cells (i.e. 40 km) behind the modelled present-day position. For the Greve model, this distance is one grid cell (i.e. 40 km, the Greve forcing). Apparent differences exist between "Greve1" and "Greve2". While the ice-sheet margin is fixed in position after 8 ka BP in "Greve1", the output of "Greve2" shows a more dynamic behaviour of the ice sheet after the Orkendalen period (Fig. 4).

**DISCUSSION**

How can we explain the observed differences between the models and the geology and also between the three models in terms of model characteristics? In this discussion we concentrate on the topics described above.

The contrast between the poorly modelled ice-margin positions in the coastal area and the reasonably well modelled deglaciation pattern on land may be sought in the period prior to 12–10 ka BP. The modelled ice sheet does not reach the present-day coastline in time and continued deglaciation is postponed until 11 000 calendar years BP by climate cooling related to the end of the Bolling/Allerod and Younger Dryas. In the near-coastal areas, large uncertainties exist in the geological model. Nevertheless, if we assume that the geological record is correct, we may conclude that modelled ice-margin positions for the Fabre and Huybrechts models during the period 15–12 ka BP are flawed. This is probably related to an incomplete treatment of the influence of sea level on ice-marginal ice wastage. This may occur because isostatic effects that cause a regional deviation in
relative sea level in West Greenland from the global sea-level record of Chappell and Shackleton (1986) used by
"Huybrechts" are not sufficiently modelled, or as a result of
the internal model physics, i.e. the calving method
adopted. The effect of sea-level forcing can be illustrated
by the time at which modelled ice sheets reach the
present-day coastline. Greve's model ignores sea-level change and has a difference of 8 ka with the geological
record. Using present-day sea level does not allow the
ice sheet to expand on the present-day shelf. When sea-level
changes are included, the ice sheet can expand on to
the shelf. The deglaciation on the present-day shelf is sensitive
to the prescription of sea level and ice wastage. These two
factors determine whether the age of modelled margins in
the coastal area is in agreement with the geological
record. Considering the time at which the ice sheet
reaches the present-day land area, the conceptual
treatment of sea level used by "Fabre" works out well.
The age difference between models and geology which
develops due to the incorrect treatment of sea level on
ice-marginal ice wastage has consequences for the modelled
ice-margin position in the land area. This may explain the
postponed retreat in the coastal area for the Huybrechts
model. The Fabre model is slightly ahead of the
general record in the on-land area. The difference
between Fabre and Huybrechts is probably related to
differences in the ice-core records used.

The deglaciation rate is reasonably modelled by all
three models. Part of this success may be attributed to the
absence of ice streams in the area. Although large fjords
exist, ice streams probably only developed in the near-
coastal area. Ice streams are an extreme case of ice motion
determined by basal processes. The role of basal sliding in
determining ice-margin positions is probably relatively
unimportant in the part of the West Greenland area
examined in this study. However, a basal-sliding module
may be a crucial prerequisite to model relatively large
fluctuations (retreat and advance) such as occurred
around the Holocene climatic optimum or during the
cold spell around 8.0 ka ago. In the marginal area, mass
flow via basal sliding makes a rapid advance possible,
with moderate ice thicknesses in marginal grid cells,
relative to the case that ice thickening in the marginal
grid cells must originate by ice creep only. The moderate
ice thicknesses in marginal grid cells can subsequently
easily ablate. Therefore, basal sliding enables the ice sheet
to become more dynamic in its response to climatic
forcing.

The synchrony in deglaciation rate between the
models and the geology further implies that the increased
ablation during deglaciation predicted by the parameter-
ization of surface melt is reasonably correct. An important
factor which determines this success is the quality of the
forcing, i.e. the ice-core records used.

After the Orkendalen period, the ice margin retreated
behind its present position. Near Jacobshavn, Weidick
and others (1991) found evidence for a substantial retreat
within this period known as the hypsithermal or Holocene
climatic optimum. The recession is conceived as "natures
own greenhouse experiment" (Weidick and others, 1991)
and provides the opportunity to judge the performance of
models which are used to depict future changes of the
Greenland ice sheet.

The Greve2 and Huybrechts models generate a retreat
behind the present-day ice margin. Differences between
Greve2 and Huybrechts are probably related to the origin
of the forcing (GRIP versus Dye 3 (Camp Century). The
GRIP record may be less valuable for the Holocene,
because it is not corrected for flow effects. The
representation of the signal of the Holocene climatic
optimum is also too weak in the GRIP record (paper in
preparation by S. Johnson and others).

A remarkable phenomena is that model Greve1,
which is led by a smoothed version of the GRIP curve
with 2000 year averages, does not produce a dynamic,
fluctuating ice sheet around the Holocene climatic
optimum. When the unsmoothed GRIP data are used
(in Greve2) the ice sheet advances to its present-day
position following a retreat of one grid point. The GRIP
record, used to force surface temperature and accumulation,
does not show a large climatic change throughout the
Holocene (Dansgaard and others, 1993). The already
weak climatic signal related to the Holocene climatic
optimum is eliminated by smoothing.

An additional effect may be the impact of short-term
decadal) fluctuations in climatic variables on the
position of the ice margin. Many short-term fluctuations
in δ18O can be observed in the GRIP record during the
Holocene. These fluctuations have a duration of decades
and a magnitude of about 0.5‰, and some have a
periodic nature (Tatenhove, 1995). The impact of these
fluctuations is probably not detectable within the spatial
resolution of the ice-sheet models discussed, because the
associated fluctuations of the ice margin have a magni-
tude of several hundreds of metres up to several
kilometres. Experiments with a surface-energy balance
model of the Greenland ice sheet showed that random
variations, with a frequency of 2 years, cause an increase
of ablation by 10% under constant climatic conditions
(Wal and Oerlemans, 1994).

Therefore, it is suggested that the elimination of the
weak climatic signal by smoothing, and possibly the effect
of decadal fluctuations in ablation and accumulation rate,
explain the difference between models Greve1 and
Greve2. It clearly shows the sensitivity of modelled ice
margins for trends in forcing with long-term or short-term
signals. Models which are going to be used to evaluate the
future geometry of the ice sheet should include the effect
of short-term variability in climatic parameters.

CONCLUSIONS

All three models for which results are discussed belong to
the same class of model, i.e. they take account of the
coupling of temperature and velocity fields on a three-
dimensional network. Despite this similarity, distinct
differences exist in modelled positions of the ice margin
in West Greenland from the Last Glacial Maximum to
the present. Disagreement between the geological record
and modelled margins in the near-coastal area can be
explained by inadequate modelling of marginal ablation
in a tide-water environment, either by inadequate forcing
functions for sea-level fluctuations or by an insufficient
parameterization of calving.

The parameterization of surface melt and surface
temperature based on the GRIP or Dye 3/Camp Century data (Dansgaard and others, 1993; Huybrechts, 1994) is capable of simulating the pattern of deglaciation rather well. This success may partly be the result of the apparently subordinate role of basal processes (sliding, bed deformation) in the study area in West Greenland.

The Greve and Huybrechts models show a smaller ice sheet than at present during the Holocene climatic optimum. In the case of the Greve model, this smaller ice sheet only occurs when an unsmoothed forcing is used which correctly treats the climatic signal related to the Holocene climatic optimum.

During the Holocene, short-term variations in ablation as depicted by ice-core records will result in margin fluctuation of several kilometres at best. Such fluctuations are not detectable using present-day ice-sheet models. However, the length of the ice margin in West Greenland which is sensitive to fluctuations during the Holocene climatic optimum is about 600 km (Weidick, 1993). Fluctuations of the ice margin with a magnitude of several kilometres may therefore cause global sea-level changes of several tens of millimetres. This underlines the importance of a more refined spatial resolution of models and the incorporation of short-term variations in climatic variables.

REFERENCES


