QUATERNARY RESEARCH 44, 317–327 (1995)

Glacial-Geological/Geomorphological Research in West Greenland Used to Test an Ice-Sheet Model

FRANK G. M. VAN TATENHOVE AND JAAP J. M. VAN DER MEER

Fysisch Geografisch en Bodemkundig Laboratorium, Universiteit van Amsterdam, Nieuwe Prinsengracht 130, 1018 VZ Amsterdam, The Netherlands

AND

PHILIPPE HUYBRECHTS

Alfred Wegener Institut für Polar-und Meeresforschung, Columbusstrasse, D-27515 Bremerhaven, Germany

Received November 17, 1994

Ice sheet modeling is an essential tool for estimating the effect of climate change on the Greenland ice sheet. 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. A framework for the analysis of glacial geological data to test ice sheet models is illustrated by a case study in west Greenland. A geological scenario of ice margin positions since the last glacial maximum is based on a review of existing literature and new datings of moraine systems. Ages of moraine systems and associated accuracy ranges are interpolated to a 1 × 1 km grid over an area of about 57,500 km^2. Resampling to a 20 × 20 km grid on an ice sheet model coordinates permits a quantitative comparison with modeled ice marginal positions. In view of the uncertainties in the geological scenario, with moraine system ages having an absolute uncertainty of 700 to 3300 cal yr, the Greenland ice sheet model of Huybrechts provides a reasonable simulation of the deglaciation pattern in central west Greenland. Modeled timing of the position of the ice margin generally precedes the geological record by 900 yr. The difference between geology and model is large for areas without proper geological information and small, but still about 500–950 yr, for well-dated moraine systems. Model deficiencies in west Greenland are probably related to the forcing function driving ablation estimates, and especially the forcing function for sea level and the description of calving processes should be reviewed critically. The simplified topography used by the model could also induce errors. Although topography used in models neglects possible important features such as ice streams, we do not find any significant trend between differences and relief amplitude. Missing moraines on the southern part of the shelf and absence of geological information from beneath the present ice sheet affect the character and quality of the geological scenario.

INTRODUCTION

This study is part of a larger project that aims to establish more accurate predictions concerning the contribution of the Greenland ice sheet to global sea level rise within the next few centuries (Oerlemans and Vugts, 1993). In 1990, a glacial-geological research program began with the purpose of providing ice sheet modelers with data on the Holocene retreat and fluctuations of the western margin of the Greenland Ice Sheet at approximately 67°N. The geological model will be used to compare geological observations with the predictions of glaciological models. The performance of models should be evaluated for as many situations as possible. Every time a model passes the test, more confidence may be given to model results for periods without control options, such as the future. The confirmation of models is inherently partial because they are fed by input parameters for which there is imperfect knowledge (Oreskes *et al.*, 1994). Also the loss of information introduced by the use of continuum mechanics at a scale lower than the average scale makes the description of reality by models fragmentary. The concept of confirmation of ice sheet models using geological information, or more-specific ice margin positions, is well accepted (i.e., Paterson, 1981; Haeberli and Schlüchter, 1987; Payne and Sugden, 1990). Although the concept is simple, in practice there are several discrepancies between geological "reality" and ice sheet model "reality" (Payne and Sugden, 1987). This paper will explore some of these discrepancies and highlight them by using the west Greenland transect as a case study.

For comparison between glacial geology and a numerical model, only geological information directly related to the model output can be used. Glacial-geological information generally consists of an array of different types of sedimentological and morphological data on scales ranging from 10^{-6} m (e.g., deformational structures in till seen in thin sections; Van der Meer, 1993) to 10^4 m (linear structures visible in Landsat images; Clark, 1993). It is often unclear in what way geological data are related to ice sheet models. This is partly due to the large uncertainties involved in a quantitative interpretation of sedimentological and morphological phenomena. On the other hand, the perception of glacial geologists did not encourage a look at geological data from a modelers viewpoint.

The aims of this paper are (1) to review the possibilities and limitations of glacial geology and, more specifically, the use of ice-marginal features to evaluate ice sheet models; (2) to pro-

0033-5894/95 $12.00

vide a methodology to enable comparison between ice sheet model results and geological evidence of ice margin positions; and (3) to evaluate an existing Greenland ice sheet model on the basis of the geological evidence and methodology of comparison.

This paper does not deal explicitly with the characteristics of ice sheet models. A technical discussion of ice sheet models can be found in Oerlemans and Van der Veen (1984), Huybrechts (1990), and Hindmarsh (1993). The study area in west Greenland used in this study stretches from the continental shelf break to the present ice sheet margin (Fig. 1).

FRAMEWORK FOR THE ANALYSIS OF GLACIAL GEOLOGICAL DATA TO TEST ICE SHEET MODELS

Introduction

Glacial-geological data can be grouped according to their relationship with ice sheet models (Table 1). Data used to force an ice sheet model are *input data.* Ice sheet models require forcing functions for mass balance and surface temperature. The use of paleoclimatological data derived from glacial-geological studies as input for ice sheet models is not considered because better sources are available, such as ice-core records, detailed records of ocean sediments, or records of carbonate precipitation. However, reliable geology may function as "input" and the ice sheet configuration required to produce this geology may be assigned as "output." Although we recognize the application of geology in this respect, this study has the opposite aim, to use geology to test model output. Another essential input variable is topography.

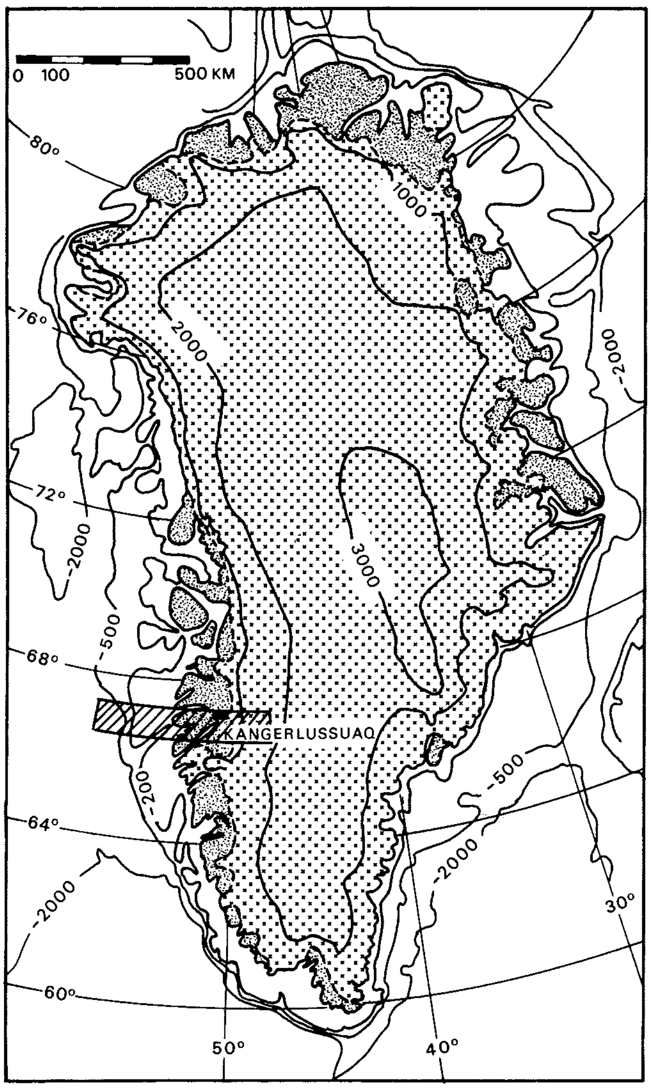

FIG. 1. Map of Greenland showing the location of the area within the rectangle near Kangerlussuaq displayed in Figures 2, 3, and 4.

Geological evidence for certain conditions at the glacier base (bed material, subglacially deformed sediments, patterns of subglacially formed lineations), evidence for conditions at the margin (land or water based, cold or temperate), and topographical features with influence on the local mass balance or flow conditions (fjords, or topography leading to formations of ice-dammed lakes) are not related to model output, but they can be used to explain inconsistencies in ice sheet model results. These type of data are called *ice-sheet-related data.* Nearly all glacial-geological data reported in the literature are of the latter type.

Processes responsible for subglacial deposition or erosion are not taken into account in present Greenland ice sheet models. We therefore do not attempt to compare records of subglacial deposits or erosional features with ice sheet models. However, we would like to stress the importance of such records in evaluating inconsistencies in ice sheet models or in reminding modelers of the existence of a complex "real" world. In this paper we will concentrate on *test data.*

Test data

Elements which are output of an ice sheet model and which can be traced in the field are called *test data.* Normal output of an ice sheet model is the position of the ice margin, ice volume, ice thickness, basal shear stress, basal temperature, and changes of these elements through time (Table 1). The elements which can be used to relate geology with ice sheet models are limited by scale. On the scale of present Greenland ice sheet models (grid size 20 km), the position of the ice margin is the element which can be used to test the model from surficial geology. The glacial-geological features used for this purpose are frontal ice-marginal features (Table 1). Estimating ice thickness from geomorphological data is only possible if lateral depositional features exist in an outlet glacier setting. A typical scale for such an environment is 1–10 km, which will be averaged out at larger scales. Assessing ice thickness without information on lateral features is possible using preconsolidation test of subglacial sediments (Boulton and Dobbie, 1993) or using isostatic rebound as a measure of ice load during maximum glaciation conditions (Peltier, 1994). The reconstruction of ice sheet volumes is not possible solely on the basis of ice margin positions, as estimates of former ice sheet profiles are also needed. Table 1 summarizes the possibilities

TABLE 1
Glacial Geological Data at Two Scales to Test Ice Sheet Models and the Geomorphological Objects Used in the Greenland Case Study

Model output	Glacial geological features to constrain model		Greenland case study	
	Ice sheet scale	Outlet glacier scale	Ice sheet scale	Outlet glacier scale
Position ice margin	Frontal ice-marginal features	Frontal ice-marginal features	Moraine systems	Moraines
Ice thickness	Consolidation tests, isostatic rebound	Lateral ice-marginal features	—	Lateral moraines and estimated valley infill
Ice sheet volume	—[a]	—[a]	—[a]	—[a]
Flow direction	Flow parallel erosive and depositional features	Flow parallel erosive and depositional features	—[c]	—[c]
Basal temperature	Subglacial depositional and/or erosional features, deep ground temperatures[b]	Subglacial depositional and/or erosional features, deep ground temperatures[b]	—[d]	—[d]

[a] Ice sheet volume can not be reconstructed from glacial-geologial data only.
[b] Glacial-geological data can only provide qualitative constraints at present.
[c] Flow direction in west Greenland is obvious; longitudinal depositional features have not been found or reported in literature. Striae directions are consistent towards west.
[d] In west Greenland limited information is available on subglacial sediments (mainly because of permafrost).

of glacial-geological data at two scales to test ice sheet models and the geomorphological objects used in the Greenland case study. If the glacial-geological features are dated (absolute or relative), a chronology of the features can be established and compared with model output.

Limitations and Uncertainties of Glacial Geological Test Data

For the west Greenland case study presented in this paper, we only make use of the position of the ice sheet margin in time as deduced from morphological evidence (frontal moraines) tied in time by ^{14}C dates. Despite the fact that moraines are real features, their interpretation with respect to processes of formation and position within a stratigraphic record is open to uncertainties. The resulting uncertainty is difficult to quantify, but should nevertheless be considered when comparing geological data with ice sheet models. Uncertainties are introduced by: (1) The ^{14}C date itself, which is related to material sampled and ^{14}C measuring errors. (2) The relation between dated material and a moraine system. Because moraines are generally not dated directly, a relation must be assumed between the geomorphological and sedimentological characteristics of the ^{14}C sample location and the associated moraine system. (3) Inaccurate topographic maps. Standard topographic maps of the ice-free areas of west Greenland near 67°N latitude, on a scale 1:250,000, have an error in position of at least 1–2 km, while the error in altitude is ±50 m in the coastal area and can be more than 200 m near the ice sheet margin (A. Nielsen and Kort-og Matrikelstyrelsen, personal communication, 1995). Mapping of ice margin position in west Greenland is therefore only possible with a positional error of 1–2 km (although present satellite positioning systems enable much more accurate mapping, the available geological information uses the inaccurate 1:250,000 topographic map). The degree to which positional errors affect the quality of geological data used to test ice sheet models depends on the scale of the model. An error in position of 1–2 km is less important for an ice sheet model with a 20-km grid than for an outlet glacier model (typical grid size 0.1–1 km). Subglacial topography of the present Greenland ice sheet, estimated from radio-echo soundings, is accurate within about 140 m (Letréguilly *et al.,* 1991). (4) Scaling-up of nonadditive geological properties. Observations of geological structures are typically on the order of 10^{-1} m (borehole), 10^{0} m (sections), or 10^{2}–10^{3} m (morphological features). The geological scale is therefore 10–1000 times smaller than the output scale of current ice sheet models for Greenland (Huybrechts, 1994). Geology must be transfered to model scale to allow comparison. Geology (i.e., moraine age) at the model scale is an interpretation of reality (moraines) which do not exist at model scale (i.e., there are now moraines of 20 × 20 km). The continuum mechanics used in these ice sheet models do not include finer-scale processes responsible for nearly all observed geological features.

Ice sheet models produce ice margin positions at every time step. However, the geological record will be an incomplete and discontinuous reconstruction of former ice sheet extent. Not all ice margin positions leave evidence, and syn- or postdepositional erosion can destroy evidence of ice-marginal deposits. Evidence of earlier, less-extensive advances is destroyed by more recent ones.

We do not use information on the genesis and size of moraines because observed morphology and sediments can have multiple interpretations. Lack of knowledge of the relationships between morphology and responsible geomorphological and glaciological processes limits the possibility of extracting quantitative information from geomorphological features to relate to ice sheet dynamics. We also do not assign different degrees of relevance to moraine systems, except for limiting our study to moraine systems traceable over large distances (10s of km). Regional significance is attached to moraines if they are traceable over large areas and continue without being interrupted by topography, such as valleys (e.g., Ten Brink, 1975).

A further subdivision of ice-marginal features by their extent (local, regional, or global), their position within sequences of morainic ridges, or size and genesis of the moraines, is not appropriate within the light of the available data. An exception is made for the geological evidence of maximum glacier extent, identified over large distances in many parts of the world, during several periods in the geological record. Such a position within a moraine sequence is regarded as highly relevant because it reflects the transition from periods of positive mass balance to periods of negative mass balance.

THE WEST GREENLAND CASE STUDY

Glacial Geology of the Study Area

Figure 2A illustrates the topography of the area between 66°30′–67°30′N and 49°25′–58°25′W. The continental shelf has a width of about 100 km. The shelf is divided into individual shallow bars (depth <50 m) by a series of deep transverse channels which reach depths of >400 m. One of these channels can be found southwest of Sisimiut (Brett and Zarudzki, 1979). The physiography of the terrestrial ice-free area is characterized by three zones: the coastal zone, a flat area with an altitude less than 200 m; the mountain zone, with altitudes up to 1500 m and a high relief amplitude caused by the incision of fjords; and the inland zone, a hilly landscape with mountains up to 600 m and a more gentle relief. The fjords in the area typically have depths of 200–600 m and a sill of variable width on their seaward side. The position of moraine systems is taken from Weidick (1968), Ten Brink and Weidick (1974), Ten Brink (1975), Brett and Zarudzki (1979), Funder (1989), and Kelly (1985), and mapping associated with the present study. A study of the aerial photographs of the area provided a check on the published data. The outlines of moraine systems have been digitized and coordinates converted to UTM to facilitate calculations involving distances.

Moraine systems are divided into four groups depending on the way ages are determined. Group A and D contain moraine systems without ^{14}C age control. The relation between relative sea levels, associated with deposits containing ^{14}C-datable organic matter, i.e., marine shells, and moraine systems, is used to provide ages for group B. Group C is dated by terrestrial organic matter found between morainic ridges. Although some ^{14}C dates can be used to constrain the age of group D moraines, this group lacks the evidence of position, because it is now covered by the ice sheet.

The ages of moraine systems are given in Table 2 and are based on the work of Ten Brink and Weidick for Group B (Weidick, 1972; Ten Brink and Weidick, 1974; and Ten Brink, 1975). The ages for the Umîvît/Keglen and Ørkendalen systems are based on studies by Van Tatenhove, (1995).

The ranges of moraine system age in Table 2 include the time period encompassed by the moraine system, the error related to the ^{14}C-dating itself, and the uncertainty due to inconclusive relationships between a ^{14}C date and a moraine system. For the Hellefisk and Sisimiut moraines, no ^{14}C dates are available within the research area and the assigned dates are based on studies by Van Tatenhove (1995). Moraine systems with a limiting ^{14}C date close to the former ice margin (Fjord, Umîvît/Keglen) have a relatively small age range. When using terrestrial organic matters the dates obtained generally place a minimum age on the associated moraine system. However, dates of marine and terrestrial organic matter are difficult to compare without calibration. Results of ice sheet models are generally expressed in calendar years, which provides an additional need for calibration. The calibration program of Stuiver and Reimer (1993) was used for calibration. Calibration introduces an extra uncertainty in time as expressed in column 4 of Table 2.

Marine shells used for dating are found in fjords and valleys. The associated moraine systems often have north–south extents of 10 to 100 km. That the continuity in morphology reflects synchrony is doubtful. The uncertainty in moraine system age therefore increases with distance from ^{14}C samples. This uncertainty is not taken into account in the present analysis.

Geological Scenario

The digitized moraine systems (Fig. 2B) in a grid of 1×1 km were assigned a date and uncertainty in calendar years (δ age). δ age is the uncertainty range of moraine system age due to the geological uncertainties (^{14}C dating, relation of ^{14}C date to moraine system, and the time span encompassed by the geographical extent of the moraine system). Before scaling up to a 20×20 km grid, we interpolated moraine age and the accompanied uncertainty range to those 1×1 km cells that are without geological information (Figs. 2C and 2D). This provides the opportunity to use additional expertise on deglaciation dynamics since the areas in between moraine systems are likely to have intermediate ages. Interpolation adds new uncertainties to moraine age. Therefore, the uncertainty for interpolated grid cells is larger than for cells containing moraines. Moraine ages (δ age) are interpolated using an inverse distance routine with power 1 to gridcells with an altitude >–175 m. The latter restriction is set to avoid assigning moraine ages below the shelf break. The uncertainty in age for interpolated cells is calculated from the δ age of nearby moraine systems assuming that the interpolated value must be proportional to distance and that the uncertainties in dates of moraines systems are independent and random. Dates from surrounding moraine systems place an absolute constraint on the interpolated age and uncertainty range. In order to obtain an age value for the ice sheet models' 20×20 km grid cells around the models' grid point coordinates, 400 cells of 1×1 km have to be averaged around the model grid point coordinates (Fig. 3A). At present we assign a single moraine age to each grid cell. This introduces a scale-related uncertainty. If, within a 20×20 km grid of an ice sheet model, two ice margins are accurately dated, the average

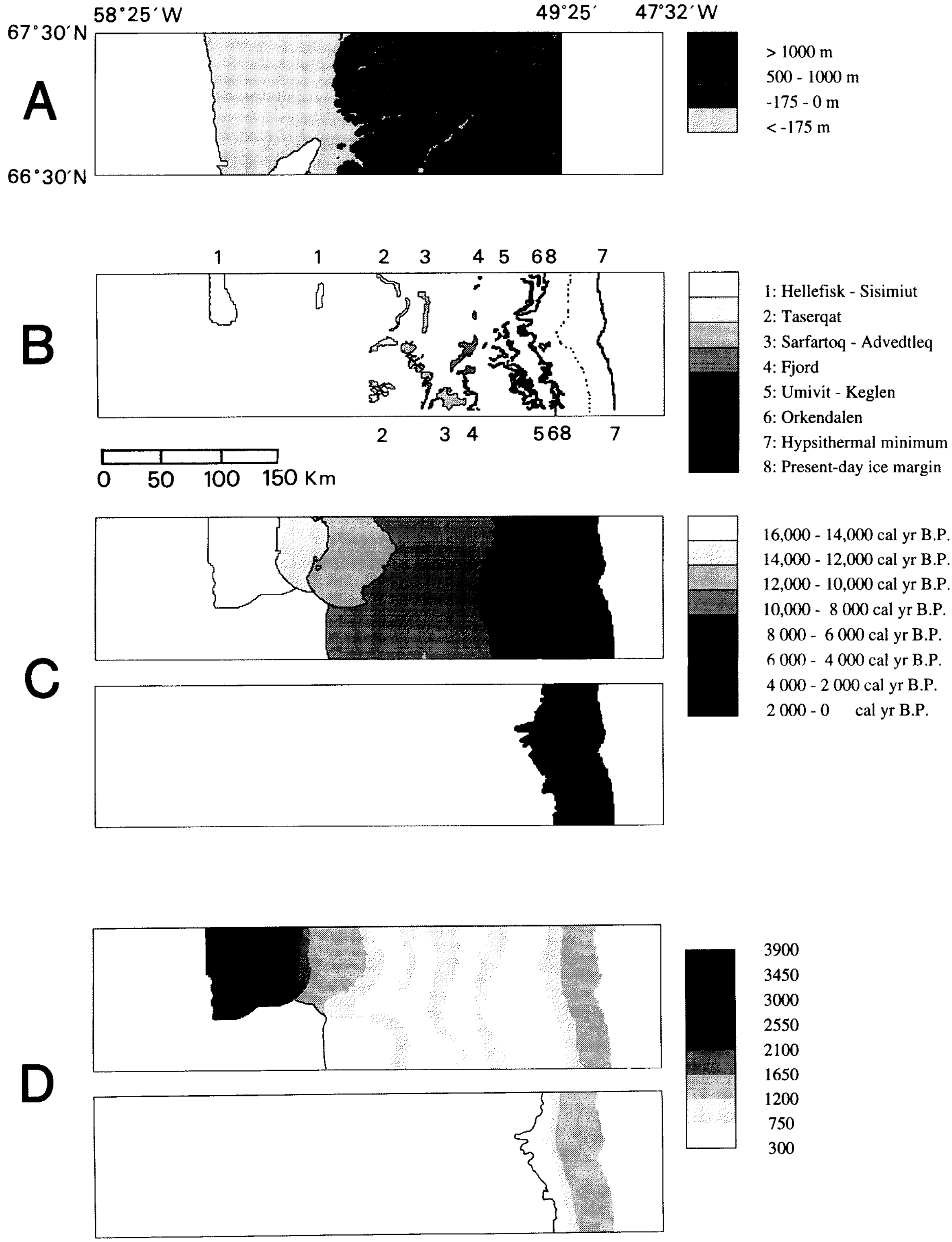

FIG. 2. (A) Topography of the area between 66°30′N–67°30′N and 49°25′W–58°25′W. (B) Moraine systems in central-west Greenland. See text for references. (C) Interpolated moraine age at a grid of 1 × 1 km. (D) Interpolated uncertainty of moraine age (3 δ age in cal yr) at grid 1 × 1 km.

deglaciation date for the grid will be less accurate than the individually dated ice margin positions. The sample standard deviation of these 400 cells (σ_s) is used as an estimate of the uncertainty involved in scaling up nonadditive geological properties (moraines). The lower limit of δt_m, is given by δ age and neglects scale-related effects. The upper limit gives equal weight to scale and geological related uncertainties and is given by $\sqrt{\delta\ \mathrm{age}^2 + (3\sigma_s)^2}$.

TABLE 2
Deglaciation Chronology of West Greenland

Group	Moraine formation stage	Age (^{14}C yr B.P.) (± absolute range)	Age (cal yr B.P.) (± absolute range)	Absolute range (cal ^{14}C yr B.P.)
A	Hellefisk		16,000 ± 3000	19,000–13,000
	Sisimiut		12,300 ± 1500	13,800–10,800
	Taserqat		9900 ± 600	10,500–9300
B	Sarfartôq-Advedtleq	8800 ± 300	9200 ± 600	9800–8600
	Fjord	8300 ± 300	8500 ± 600	9100–7900
	Umîvît-Keglen	7000 ± 500	7300 ± 600	7900–6700
C	Ørkendalen	5900 ± 300	6800 ± 300	7100–6500
D	Minimum position Hypsithermal	4000 ± 900 (?)	4000 ± 900[a]	4900–3100
	End advance Neoglacial	younger than 625 ^{14}C yr B.P.	AD 1750 ± 100[b] 200 ± 30	260–110

Note. Division in groups is based on the availability and type of age determination. (A) No ^{14}C dates available in off-shore area; (B) ages based on radiocarbon-dated marine shells; (C) ages based on radiocarbon-dated terrestrial material; and (D) no ^{14}C dates available in area presently covered by ice sheet.

[a] Geological model is developed assuming that the Hypsithermal minimum position of the ice sheet is 50 km east of the Ørkendalen moraine system.

[b] Weidick (1985).

The final uncertainty of the age of former ice margin positions, δt_m (Fig. 3B), is assumed to be the arithmetical mean of the two limits:

$$\delta t_{\mathrm{m}} = \frac{1}{2}\,\delta\ \mathrm{age} + \frac{1}{2}\sqrt{\delta\ \mathrm{age}^2 + (3\sigma_s)^2}. \qquad (1)$$

Because model output and geological reconstruction are on different projections, the final comparison between geology and the numerical model is made using the precise coordinates of ice sheet model gridpoints after transformation to UTM. The grid point coordinate is then the center of a 20 × 20 km area on which moraine age and uncertainty are resampled from the 1 × 1 km grid. In Figure 3C, δt_m is given for four characteristic regions: the offshore area without ^{14}C-dated moraine systems (group A), the moraine systems dated with marine shells (group B), the Ørkendalen system (Group C, dated with terrestrial organic matter), and the presumed moraine system behind the present ice sheet margin (Group D). The relative importance of the two components in δt_m is different for each region. In the offshore area δt_m is large and 95% caused by the absence of ^{14}C dates. The marine and terrestrial dated moraine systems have a δt_m of similar magnitude (≈700 yr). For marine samples, the uncertainty introduced by calibration is about a third of δt_m, mainly because of the reservoir effect. For terrestrial dated systems, calibration does not introduce large uncertainties (4% of δt_m). The well-dated Ørkendalen system is within 1 gridcell of other moraine systems and therefore the effect of scale is relatively large (32% of δt_m).

The final deglaciation scenario on a 20 × 20 km grid is given in Figures 3A and 2.3 B, the geological "reality" which can be compared with deglaciation dates from models.

The method to retrieve a geological scenario based on interpolated moraine ages as proposed in this paper has several advantages:

It is a systematic, quantitative measure of geology on a scale and typology which can be compared with ice sheet models. One of the main results of this study is to promote the use of geological features that can be related directly to ice sheet model output, rather than in a diffuse, descriptive way.

It includes a measure of uncertainty.

It is flexible. New findings (e.g., additional ^{14}C ages) are easily incorporated to produce an up-to-date geological scenario.

It can also be used to evaluate future research in simulating the effect of better geological information on the geological scenario and its uncertainties.

Glaciological Model

The model output discussed in this paper originates from a model run presented by Huybrechts (1994). In that study, a reconstruction was made of the ice sheet's history over the last glacial–interglacial cycle, with special emphasis on the ice sheet's evolution between the last glacial maximum and the present. The ice sheet model covers all of Greenland and distinguishes between mass flow resulting from internal deformation and basal sliding. The softness parameter of ice, which determines the rate of deformation, is a function of both ice temperature and age of the ice. For that reason, the three-dimensional temperature and velocity fields are calculated in the coupled mode and there is a rigorous tracking of particle trajectories. Isostatic compensation in response to the changing ice loading is treated by assuming local hydrostatic equilibrium and a viscous asthenosphere. The horizontal resolution of the model is 20 km and there are 26 layers in the vertical. An overview of the fundamental mathematical equations governing the model was given by Huybrechts *et al.* (1991).

The main output of the model is the 3 D time-dependent ice-sheet geometry, which is freely generated in response to prescribed environmental conditions. These are surface temperature, mass balance, and the location of a coastline, beyond which the ice, if any, is lost to the ocean. The external forcing

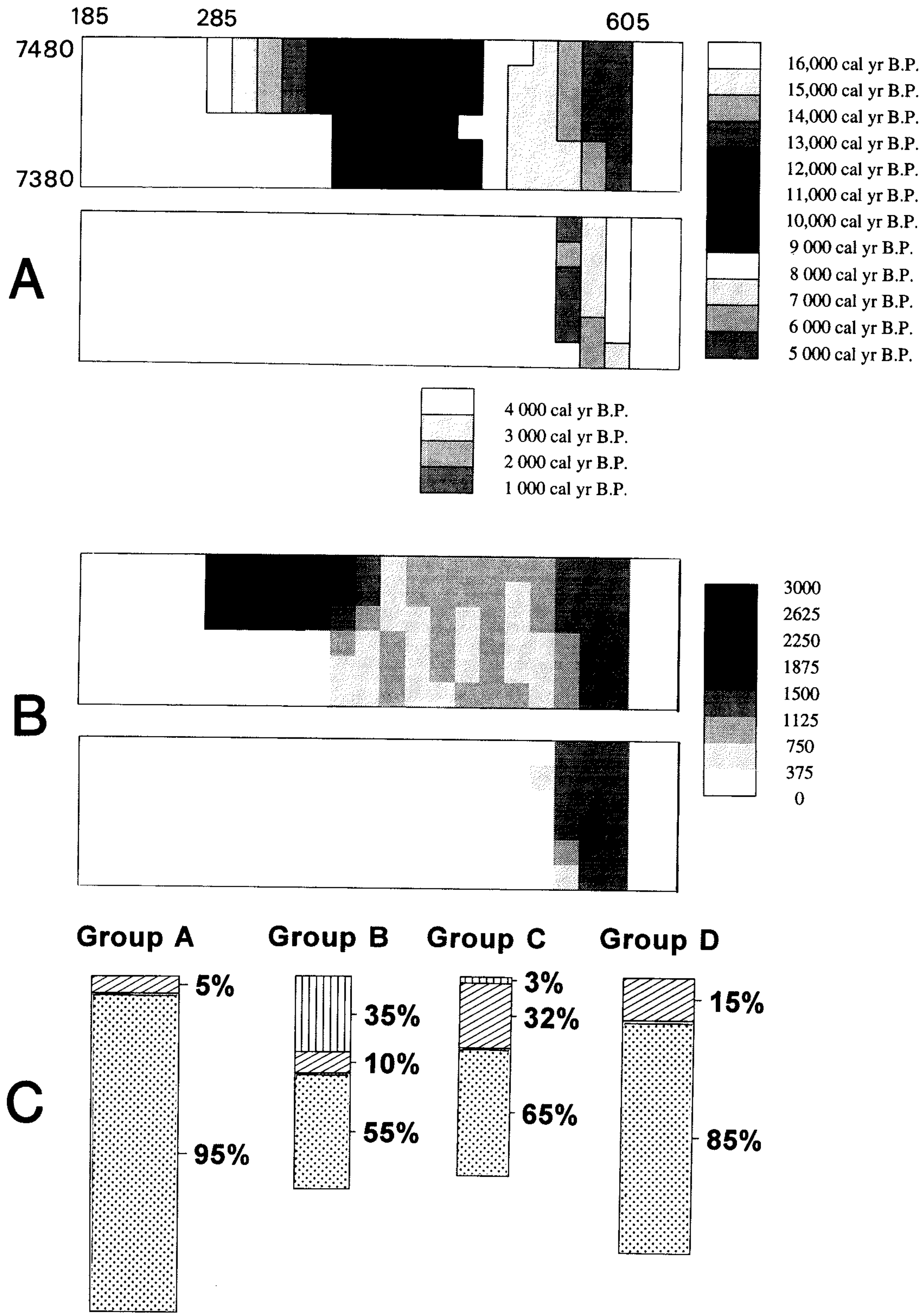

FIG. 3. (A) Moraine age at 20 × 20 km grid. (B) Absolute range of uncertainty in moraine age at 20 × 20 km grid (δt_m in cal yr). (C) The contribution of geology, scale and calibration in δt_m for four characteristic regions, the offshore area without ^{14}C-dated moraine systems (Group A), the moraine systems dated with marine shells (Group B), the Ørkendalen system (Group C, dated with terrestrial organic material), and the presumed moraine system behind the present ice sheet margin (Group D). Dots, part of δt_m caused by geological uncertainties; diagonal lines, part of δt_m caused by scale; vertical lines, part of δt_m caused by calibration.

is made up by both eustatic sea level, which determines the coast line and enables the ice sheet to expand onto the shallow parts of the continental shelf during glacial periods, and a uniformly distributed background temperature change, from which the mass balance components (snow accumulation and runoff) are calculated. The melt-and-runoff model is based on the degree-day method and accounts for the daily and annual temperature cycle, a different degree-day factor for ice and snow melting and superimposed ice formation. The temperature forcing used to drive the model is representative of central Greenland conditions and was assembled from several sources. The pre-Holocene part, before 10,800 yr B.P., was derived from an ^{18}O record obtained from surface ice samples collected at the ice margin at Paakitsoq in central West Greenland (Reeh *et al.*, 1991). The temperature history since the Younger Dryas interval was collated from the Camp Century and Dye 3 oxygen isotope records (Dansgaard *et al.*, 1984; Hammer *et al.*, 1986). It is assumed that precipitation patterns resemble present patterns. Changes in the magnitude of precipitation are forced by temperature variations. Sea-level forcing is based on the New Guinea record (Chappell and Shackleton, 1986). The position of subsequent ice margins in this model run is illustrated in Figure 4 as a series of time slices spaced 1000 yr apart.

RESULTS: COMPARING GEOLOGY AND MODEL

Within the confidence limits of model and geology, there is a reasonable agreement between the simulated and observed

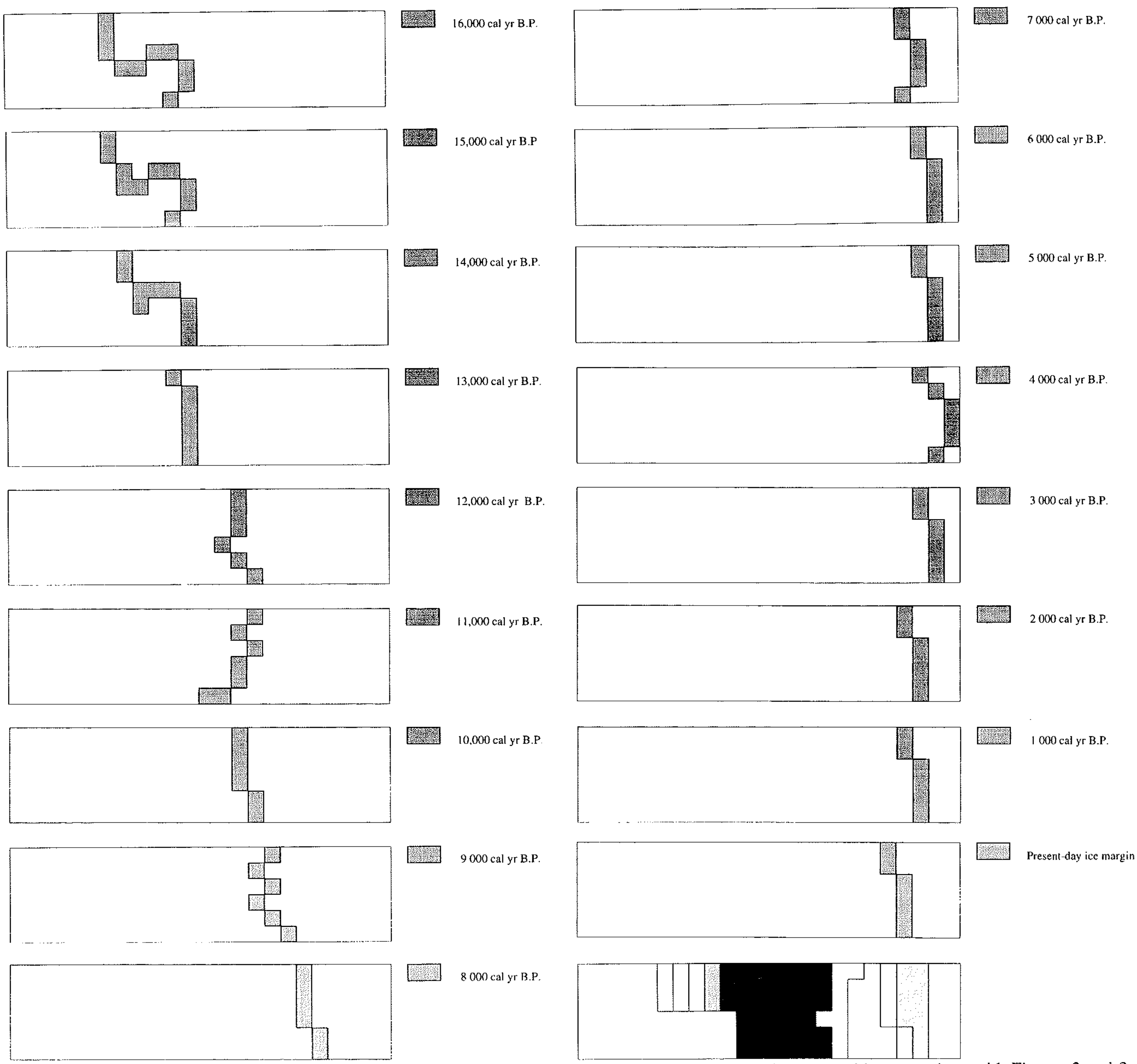

FIG. 4. Modeled ice margin position at 1000-yr intervals. Geological model in lower right panel is given to enable comparison with Figures 2 and 3.

deglaciation pattern in central west Greenland (Fig. 5). The modeled age of the position of the ice margin generally precedes the geological record (Fig. 6). The simulated present-day position of the ice margin is located one grid cell behind its actual position. On average, the difference between model and geology is about 1200 yr. This difference is large for areas without proper geological information (1900 yr) and is small, but still about 900 yr for those areas that have well-dated moraine systems (Fig. 7). Missing moraines on the southern part of the shelf (at present an area with transverse channels) directly influence the geological model. The offset of approximately 1000 yr between the modeled position of the ice margin and the geology is a remarkable result. Such a lag could conceivably be the result of model deficiencies, but can also be due to the lack of firm geological knowledge. The interpretation of the differences is complicated because a quantitative assessment of the confidence limits for the modeled ice margin positions is difficult to give. Many interdependent factors influence the model behavior.

Modelled ice-margin positions are determined by both the prescribed mass-balance (accumulation and ablation) and the ice dynamics. In the model, the maximum extent of the ice sheet during glacial periods is limited at the shelf break, where all ice not melted at the surface is assumed to calve in the ocean. This limit is taken as the coastline at a certain time step and is a function of both eustatic sea level and isostasy. During the warmer interglacial periods, on the other hand, melt rates in central west Greenland are sufficiently high to remove the ice before it can reach the coast. Errors in the calculated position of the ice margin are in that case related to deficiencies in the forcing, and in particular to the temperature signal which drives the mass-balance model and the assumption that parameters derived for present conditions are also valid in the past. Another factor of importance is the delayed response of the ice sheet to older changes in boundary conditions and bears on the response time of the model.

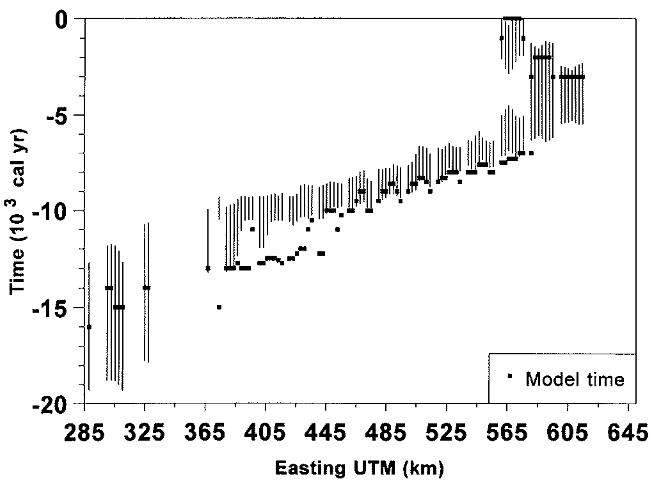

FIG. 5. Distance (in UTM) from west to east versus geological and modeled ages of ice margin position, assuming the minimum extent of the ice sheet to be 50 km behind its present position. Error bars indicate δt_m.

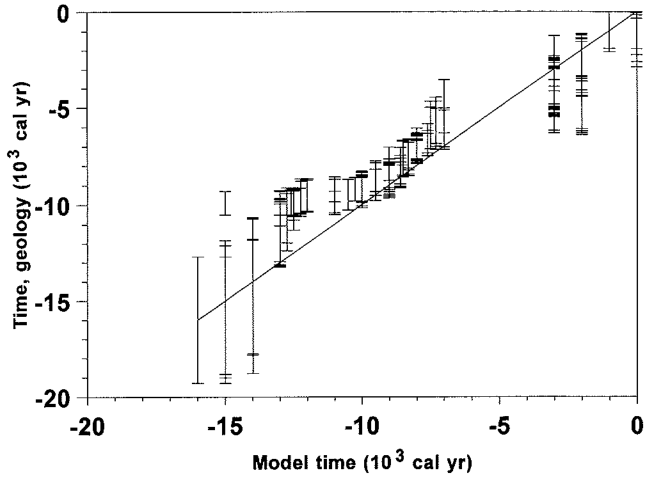

FIG. 6. Modeled time of position of the ice margin versus moraine age derived from the geological record, assuming the minimum extent of the ice sheet to be 50 km behind its present position. Error bars indicate δt_m.

From a geological point of view, the observed time lag can also be due to an underestimation of the ages of moraines at sample locations where marine shells were raised isostatically. Tidewater dynamics of glaciers within the fjord may introduce a difference in deglaciation time of several centuries between the fjord and adjacent land areas.

As illustrated in Figures 5 and 6, the modeled positions of the ice margin equals the lowest estimate of the geological scenario after about 10,000 cal yr B.P. or from 445 km UTM eastward. Before this period, the difference between geology and model is about 3000 yr (Fig. 5). Grid cells with this large offset are located in the area between the present coastline and

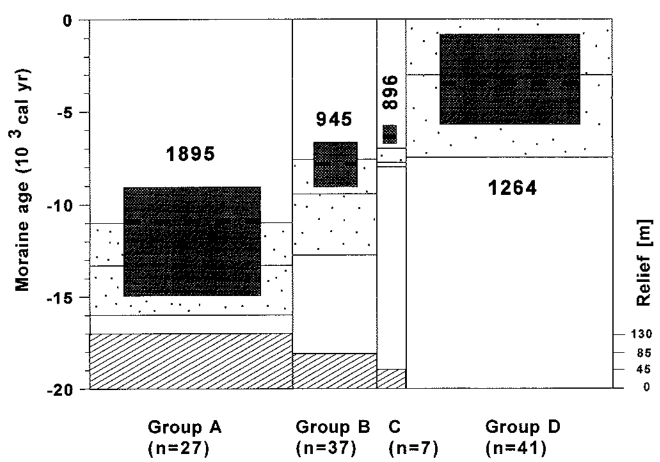

FIG. 7. Differences between modeled and geological age of ice margin position for four characteristic regions. Dashed line with gray zone is the average moraine age (gray zone expresses minimum and maximum values). Boxes with dotted signature give minimum and maximum modeled age for a certain group, where the continuous line in the center of the box give the average value. Figures in the graph are the average value of the differences between model and geology. The bars with the cross signature give the average sample standard deviation of altitude within 20 × 20 km grid cells based on the 1 × 1 km DEM of the area. No altitude values are given for Group D because nearly all grid cells are presently covered by the ice sheet.

approximately 50 km inland. The modeled deglaciation rate is similar to the geological deglaciation rate both during and after the period of large differences (Fig. 5). The explanation for the 3000 yr offset may therefore be sought in the period prior to ca. 12,000 cal yr B.P. The modeled ice sheet does not reach the present coastline in time, and continued deglaciation is postponed until 11,000 cal yr B.P. by climate cooling related to the end of the Bølling/Allerød and Younger Dryas intervals. In the near-coastal areas, large uncertainties exist in the geological model. Nevertheless, when assuming that the geological record is correct, we may conclude that modeled ice margin positions during the period 15,000–12,000 cal yr B.P. are flawed. This is probably related to an incomplete treatment of the influence of sea level on ice-marginal ice wastage. Such an incomplete treatment could be related to the isostatic effects that cause a regional deviation in relative sea level in west Greenland from the global sea level record of Chappell and Shackleton (1990), or to the internal model physics, i.e., the calving method adopted. The error works out progressively in time and causes a delayed retreat in the coastal area. The synchrony in deglaciation rate between model and geology further implies that the increased ablation during deglaciation predicted by the melt-and-runoff model is reasonably correct.

Another class of errors could have been introduced by the topography used in the model. Topography in the presently ice-free area is derived from the ETOPO5 dataset, with a resolution of 5 min of arc, and is interpolated to the model grid. Small topographical features, such as deep fjords, are smoothed out in the model input, but are important barriers to ice flow (Warren and Hulton, 1990). To investigate the effect of topographical resolution we calculated the relief amplitude within a 20 × 20 km gridcell based on 400 1 × 1 km gridcells (Fig. 7). Relief amplitude is regarded as a simple measure of the possible complexity of deglaciation dynamics. Geology within a gridcell with large relief amplitude will be more difficult to match with ice sheet model results than gridcells with a rather uniform topography. However, this crude measure of the effect of topography could not explain the pattern of differences in positions of the ice margin, mainly because it is overshadowed by the geological uncertainty (areas with a high relief amplitude have large σ_g and v.v.; Fig. 7). Even within one cluster of moraine systems, a significant trend cannot be observed (Fig. 8).

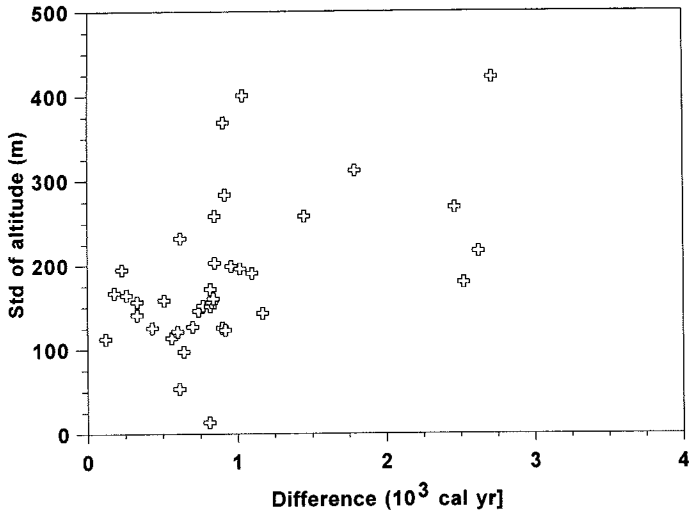

FIG. 8. Age difference between model and geology versus sample standard deviation of relief amplitude for Group B (moraine systems dated by marine shells).

CONCLUSIONS

1. Ice-marginal features are the main object obtained from glacial geological research which can be used to test ice sheet models. They can be regarded as a reliable source of information on the extent of the Greenland ice sheet during the Holocene. In spite of this, the use of ice-marginal features as independent control on the output of ice sheet models is seriously hampered by uncertainties in dating and calibration of ^{14}C dates, the unknown degree of synchrony for (apparently) continuous moraine systems, and differences in scale between present ice sheet models and moraines.

2. A geological deglaciation scenario on a 20-km grid based on current knowledge of deglaciation dynamics in central west Greenland has an uncertainty of between 100 and 3900 cal yr. The uncertainty in geology is mainly a function of the lack of dates for offshore moraines and for inferred moraines behind the present ice margin. The use of marine shells in dating moraine systems adds a significant uncertainty as a result of calibration of ^{14}C yr to calendar years.

3. In areas with several moraine systems within about 20 km, the uncertainty in moraine age of a 20 × 20 km grid cell can be significantly larger than that of individually dated moraines within the same grid cell. Although this is partly due to the approach adopted, it illustrates the importance of scale when comparing geology and ice sheet models.

4. A reconstruction of the volume of the Greenland ice sheet during Holocene deglaciation is not possible using geological evidence alone. The combination of dated ice-marginal deposits and future information on palaeoelevation of the central part of the Greenland ice sheet from ice core records will be a powerful tool in producing estimates of former ice sheet volume.

5. The Greenland ice sheet model of Huybrechts (1994) provides a reasonable simulation of the deglaciation pattern in central west Greenland within the confidence limits of model and geology. Modeled timing of the position of the ice margin generally precedes the geological record by 1000 yr. The position of the present ice margin is simulated one grid cell behind the actual position. The difference between geology and model is large for areas without proper geological information and small, but still about 500–950 yr, for well-dated moraine systems.

Differences between model and geology in the near coastal area are attributed to the incomplete treatment of the influence of sea level on wastage of the ice margin. Here, the forcing of sea level may be incorrect or the calving method adopted may be deficient. This is an example of how the comparison between the geological record and model output provides new insights in how well a modelled ice sheet resembles a real ice sheet. The synchrony in deglaciation rate between model and geology further implies that the increased ablation during deglaciation predicted by the melt-and-runoff model is reasonably correct.

The simplified topography used by the model could also induce errors in the coastal area where deep fjords may be occupied by ice streams during deglaciation. Although these ice streams may play an important role in ice sheet drainage, we find no significant trend between differences and relief.

6. Differences between model and geology are not only due to model deficiencies, but can also be attributed to lack of geological knowledge or large uncertainties in existing age estimates of moraine systems. Missing moraines in the southern part of the shelf, or absence of geological information beneath the present ice sheet directly influence the character and quality of the geological scenario.

7. Despite the stated reservations in the conclusions, we think that the present analysis might increase the use of glacial geology as an independent control on the consistency of numerical ice sheet models. It also provides several new perspectives on the Holocene deglaciation dynamics in central west Greenland, because two different, both imperfect, approaches are compared. The methodology for testing can easily be expanded to cover all of Greenland, thereby creating a more substantial test for the overall Greenland ice sheet models.

ACKNOWLEDGMENTS

This study has benefitted from discussions with Prof. J. Oerlemans (Institute for Marine and Atmospheric Research, Utrecht University). Using his ice sheet model provided the first author with the essential ''feel'' for the ins and outs of models. The Dutch National Research Programme Global Air Pollution and Climate Change provided financial funding for the first author. Detailed comments of two anonymous referees were greatly appreciated.

REFERENCES

Boulton, G. S., and Dobbie, K. E. (1993). Consolidation of sediments by glacier relations between sediment geotechnics, soft-bed glacier dynamics and subglacial ground-water flow. *Journal of Glaciology* **39,** 26–44.

Brett, C. P., and Zarudzki, E. F. K. (1979). Project Westmar. A shallow marine geophysical survey on the West Greenland shelf. *Grønlands Geologiske Undersøgelse Rapport* **87,** 28.

Chappell, J., and Shackleton, N. J. (1986). Oxygen isotopes and sea level. *Nature* **324,** 137–140.

Clark, C. D. (1993). Mega-scale glacial lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms* (18), 1–29.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N., Hammer, C. U., and Oeschger, H. (1984). North Atlantic climatic oscillations revealed by deep Greenland ice cores. *In* ''Climate Processes and Climate Sensitivity.'' *Geophysical Monograph* 29, Maurice Ewing Vol. 5, pp. 288–298. Am. Geophys. Union, Washington, DC.

Funder, S. (1989). Quaternary geology of the ice-free areas and adjacent shelves of Greenland. *In* ''Quaternary Geology of Canada and Greenland'' (R. J. Fulton, Ed.), pp. 741–792. Geological Survey of Canada.

Haeberli, W., and Schlüchter, Ch. (1987). Geological evidence to constrain modeling of the Late Pleistocene Rhonegletscher (Switzerland). The Physical Basis of Ice Sheet Modelling. *IAHS Publications* **170,** 333–346.

Hammer, C. U., Clausen, H. B., and Tauber, H. (1986). Ice-core dating of the Pleistocene/Holocene boundary applied to a calibration of the ^{14}C time scale. *Radiocarbon* **28,** 284–291.

Hindmarsh, R. C. A. (1993). Modelling the dynamics of ice sheets. *Progress in Physical Geography* **17,** 391–412.

Huybrechts, P. (1990). A 3-D model for the Antarctic ice sheet: A sensitivity study on the glacial–interglacial contrast. *Climate Dynamics,* **5,** 79–92.

Huybrechts, P. (1994). The present evolution of the Greenland ice sheet: An assessment by modelling. *Global and Planetary Change* **9,** 39–51.

Huybrechts, P., Letréguilly, A., and Reeh, N. (1991). The Greenland ice sheet and greenhouse warming. *Palaeogeography, Palaeoclimatology, Palaeoecology* **89,** 399–412.

Kelly, M. (1985). A review of the Quaternary Geology of Western Greenland. *In* ''Quaternary Environments, Eastern Canadian Arctic, Baffin Bay and Western Greenland'' (J. T. Andrews, Ed.), pp. 467–501. Allen and Unwin, Boston.

Letréguilly, A., Huybrechts, P., and Reeh, N. (1991). Steady-state characteristics of the Greenland ice sheet under different climates. *Journal of Glaciology* **37,** 149–157.

Oerlemans, J., and Van der Veen, C. J. (1984). ''Ice Sheets and Climate.'' Reidel, Dordrecht.

Oerlemans, J., and Vugts, H. F. (1993). A meterological experiment in the melting zone of the Greenland ice sheet. *Bulletin of the American Meterological Society* **74,** 355–365.

Oreskes, N., Shrader-Frechette, K., and Belitz, K. (1994). Verification, validation, and confirmation of numerical models in the earth sciences. *Science* **263,** 641–646.

Paterson, W. S. B. (1981). ''The Physics of Glaciers.'' 2nd ed. Pergamon, Elmsford, NY.

Payne, A. J., and Sugden, D. E. (1987). Problems of testing ice sheet models: A British case study. The Physical Basis of Ice Sheet Modelling. *IAHS Publications* **170,** 359–366.

Payne, A. J., and Sugden, D. E. (1990). Topography and ice sheet growth. *Earth Surface Processes and Landforms* **15,** 625–639.

Peltier, W. R. (1994). Ice Age paleotopography. *Science* **265,** 195–201.

Reeh, N., Oerter, H., Letréguilly, A., Miller, H., and Hubberten, H.-W. (1991). A new, detailed ice-age oxygen-18 record from the ice sheet margin in central West Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* **90,** 373–383.

Stuiver, M., and Reimer, P. J. (1993). Extended ^{14}C data base and revised Calib 3.0 ^{14}C age calibration program. *Radiocarbon* **35,** 215–230.

Ten Brink, N. W. (1974). Glacio-isotasy: New data from West Greenland and geophysical implications. *Geological Society of America Bulletin* **85,** 219–228.

Ten Brink, N. W. (1975). Holocene history of the Greenland Ice-Sheet based on Radiocarbon-dated moraines in West-Greenland. *Meddelelser om Grønland* **201**(4), 44.

Ten Brink, N. W., and Weidick, A. 1974. Greenland ice sheet history since the Last Glaciation. *Quaternary Research* **4,** 429–440.

Van der Meer, J. J. M. (1993). Microscopic evidence of subglacial deformation. *Quaternary Science Reviews* **12,** 553–587.

Van Tatenhove. (1995). ''The dynamics of Holocene Deglaciation in West Greenland with Emphasis on Recent Ice Marginal Processes.'' Ph.D. thesis. University of Amsterdam.

Warren, C. R., and Hulton, N. R. J. (1990). Topographic and glaciological controls on Holocene ice-sheet margin dynamics, central West Greenland. *Annals of Glaciology* **14,** 307–310.

Weidick, A. (1968). Observations on some Holocene glacier fluctuations in West Greenland. *Meddelelser om Grønland* **165**(6), 202.

Weidick, A. (1972). Holocene shore-line and glacial stages—An attempt at correlation. *Geologiske Undersøgelse Rapport* **41,** 39.

Weidick, A. (1985). Review of glacier changes in west Greenland. *Zeitschrift für Gletscherkunde und Glazialgeologie* **21,** 301–309.