Constraints on the Greenland Ice Sheet since the Last Glacial Maximum from sea-level observations and glacial-rebound models

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Abstract

Geomorphological descriptions of changes in the extent of the Greenland Ice Sheet (GIS) have been combined with glacial-isostatic-adjustment models to reproduce the sea-level history of Greenland since the Last Glacial Maximum (LGM). The contribution to past sea-level change around Greenland due to ice-load changes outside of that region has been considerable (\( \pm 10\) s of meters), while still contributing a rise of several mm yr\(^{-1}\) today. The isostatic contribution to relative sea level around Greenland from changes in the GIS is found by iteratively perturbing preliminary ice models with different LGM extents and deglaciation starting times. The resulting first-order model that provides the best agreement between observed and predicted sea level contributes 3.1 and 1.9 m water-equivalent of additional ice relative to present-day ice volumes at the LGM and Younger Dryas, respectively. The GIS in most areas does not appear to have extended far onto the continental shelf, the exceptions being southern-most Southwest Greenland and northern East Greenland, as well as at the coalescence of the Northwest Greenland and Inuitian Ice Sheets. Changes in ice thickness since the LGM were \( > 500 \) m along the present-day outer coast and \( > 1500 \) m along some parts of the present-day ice margin. The observed mid- to late-Holocene fall in sea level to below the present-day level and the subsequent transgression seen in some areas implies that the GIS retreated behind the present-day margin by distances of the order of 40 km before readvancing.

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1. Introduction

A consequence of the deglaciation of the expanded continental ice sheets after the Last Glacial Maximum (LGM, ca 18 ka BP in the radiocarbon (\(^{14}\)C) timescale, 21 ka BP in the calendar (cal) timescale) is the vertical rebound of the formerly ice-covered parts of the Earth’s surface. This resulted in the numerous raised shorelines that are observed in such areas, for example Arctic Canada, Scandinavia and the region of interest to this paper, Greenland. The Greenland Ice Sheet (GIS) was substantially larger during the LGM, as evident by the just mentioned raised shorelines as well as by recessional moraines. Studies of the ice-sheet and sea-level history of Greenland since the LGM include those dealing with the GIS itself (e.g. Ten Brink and Weidick, 1974; Kelly, 1980; Ingolfsson et al., 1994; van Tatenhove et al., 1996; Zreda et al., 1999; Bennike and Björck, 2002), local sea-level change, especially during the Holocene (e.g. Rasch and Jensen, 1997; Hjort, 1997; Long et al., 1999; Bennike and Weidick, 2001) and assessments of the isostatic contribution to sea-level change (e.g. Kelly, 1973; Ten Brink, 1974; Weidick, 1996; Bennike et al., 2002; Tarasov and Peltier, 2002).

In this paper we use observations of relative sea-level change and predictions from realistic glacial-isostatic-adjustment (GIA) models to place constraints on the primary characteristics of the GIS’s past behaviour; its extent during the LGM, the timing of deglaciation and the spatial changes in ice thickness. Such a methodology for resolving glacial histories is well established (e.g. Peltier and Andrews, 1976; Lambeck et al., 1990) and is particularly important where geomorphological constraints on the ice sheet are limited, as in the Barents-Kara Seas (e.g. Lambeck, 1995).

The interest in resolving the history of the GIS is two fold. Firstly, the GIS is the sole northern hemisphere ice sheet to have survived the Last Glacial-Interglacial Transition (LGIT), despite being in a region of vigorous
climatic change (Funder et al., 1998). Secondly, the GIS today contains around 7 m water-equivalent of ice, if evenly distributed over the present-day oceans (Le特朗éguilly et al., 1991), and is therefore an important factor when considering contemporary sea-level change (e.g. Oerlemans, 1993; Krabill et al., 2000). There is also the fact that the GIS is still reacting to past climate changes (e.g. Abe-Ouchi et al., 1994) and so resolving its former extent will assist in predicting its future behaviour.

The radiocarbon timescale is used throughout this work because most observational evidence related to sea-level change and ice-sheet evolution is constrained by radiocarbon dating. The effects of minor nonlinearities in the 14C timescale are insignificant when compared to the ice- and earth-model uncertainties.

2. Methodology

Relative sea level (Δzrel) at a time t and location φ may be expressed schematically as (Lambeck, 1993):

\[ Δz_{rel}(t, φ) = Δz_{esl}(t) + Δz_{ice}(t, φ) + Δz_{wat}(t, φ), \]

(1)

Δz_{esl} is the ice-volume equivalent sea level or simply equivalent sea level, representing the amount of ice/water exchanged between the oceans and land-based ice sheets over time (when all other effects are ignored, this value is also referred to as the eustatic sea level). Δz_{ice} and Δz_{wat} are the ice- (glacio-isostatic) and water- (hydro-isostatic) load components. Eq. (1) is a simplified description of the changes in sea level that arise from GIA, and the complete formulation used includes the interactions between the various components (e.g. Lambeck et al., 2003; Mitrovica, 2003). For example, Δz_{esl} is a function of the shape of the ocean basins, which are in turn affected by surface-load changes and by the collapse of marine-based ice sheets.

For several ice sheets, (J), and incorporating observations, Eq. (1) may be rewritten as

\[ Δz_o(t, φ) + e_o(φ, t) = Δz_{esl}(t) + \sum_{j=1}^{J} β_j [Δz_{ice}^{j}(t, φ) + Δz_{wat}^{j}(t, φ)], \]

(2)

where Δz_o is an observation’s elevation relative to present-day sea level, e_o is the correction required to relate the imperfect observation to mean sea level, Δz_{ice} and Δz_{wat} are the ice- and water-load components of the jth ice sheet and β_j is a scaling parameter for the jth ice sheet’s isostatic contribution (Lambeck et al., 1990).

Since only the GIS is considered, Eq. (2) is rewritten as

\[ Δz_o(t, φ) + e_o(φ, t) = Δz_{esl}(t) + Δz_{non}^{GIS}(t, φ) + δ_{non}^{GIS}(t, φ) + β_{gis} [Δz_{ice}^{GIS}(t, φ) + Δz_{wat}^{GIS}(t, φ)] + δ_{iso}^{GIS}(t, φ). \]

(3)

Here, Δz_{non}^{GIS} is the isostatic contribution of the non-Greenland ice sheets, i.e. those located outside of Greenland (specifically North America, Europe and Antarctica), Δz_{ice}^{GIS} and Δz_{wat}^{GIS} are the ice- and water-load contributions of the GIS, while δ_{non}^{GIS} and δ_{iso}^{GIS} are the uncertainties associated with the non-Greenland and GIS isostatic contributions. β_{gis} is the scaling factor for the GIS isostatic contribution that leads to a minimum in the variance, σ^2, defined as

\[ σ^2 = \frac{1}{(S - 1)} \sum_{s=1}^{S} \left( \frac{Δz_o^{s} - Δz_{pred}^s}{σ_o^s} \right)^2. \]

(4)

The superscript s refers to the sth observation (Δz_o^{s}) and prediction (Δz_{pred}^{s}), of a total of S data, where Δz_{pred} is the predicted change given by the right-hand side of Eq. (3) without the uncertainty parts. σ_o^s is the total uncertainty associated with the sth observation. It results from combining the correction required to relate the observation to mean sea level, e_o, which in this work is set to a constant value of 5 m, and the isostatic contributions of the non-Greenland (δ_{non}^{GIS}) and Greenland (δ_{iso}^{GIS}) ice sheets, and is written as

\[ σ_o = [e_o + (δ_{non}^{GIS})^2 + (δ_{iso}^{GIS})^2]^{1/2}. \]

(5)

σ^2 is therefore a measure of the misfit between the scaled predictions and observations. If the ice- and earth-model parameters employed are correct (i.e. give results that are within the uncertainty in the observations) and the observational uncertainties have a normal distribution, then the expected value of σ^2 is 1. However, since such conditions are rarely satisfied, a range of preliminary ice models incorporating different assumptions about the past conditions of the GIS are used, with the scaled model that best reproduces Greenland’s sea-level history selected. Because the reliability of the initial models will vary over the study area, analyses are carried out over various spatial scales, with the likelihood that different preliminary models will be preferred for different areas.

Several problems may arise with this method. Firstly, a fundamental assumption is that the former ice margins are correctly defined and that only the ice thickness needs to be modified. The method will therefore break down when this is not the case, the result sometimes being that the ice model needs to be scaled by an implausible amount to improve the fit (i.e. decrease σ^2) between the observations and predictions (e.g. Lambeck, 1995). Therefore, preliminary models with differing ice-margin locations need to be tested. Another problem concerns biases that may arise from the spatial and temporal distribution of the observations. For example, if some areas are more densely sampled than others, the resulting β_{gis} and σ^2 values will be more representative of those localities, as opposed to the entire region under consideration.
The Earth’s rheology is described by a three-layer earth model incorporating an elastic lithosphere with a thickness $H_{\text{lith}}$, an upper mantle extending to the seismic discontinuity at 670 km depth described by its viscosity $\eta_{\text{um}}$, and a lower mantle extending until the core-mantle boundary described by its viscosity $\eta_{\text{lm}}$. The mantle is assumed to be a Maxwell viscoelastic body, meaning that at low frequencies it acts as a viscous fluid while at high frequencies it responds elastically. While this is a comparatively simple description of the Earth, it is adequate for our purposes given the uncertainties involved in the ice model. Hence, the values assigned to the earth-model parameters are termed effective values.

Uncertainties associated with the modelled isostatic contribution to sea-level change arise from two sets of unknowns. The first involves our incomplete knowledge of the Earth’s rheology. The uncertainty in the predictions from this part of the problem is accommodated by using a number of earth models that include upper and lower limits of the effective values from previous GIA studies (Table 1, e.g. Nakada and Lambeck, 1989; Mitrovica, 1996; Lambeck et al., 1998; Nakada et al., 2000). Also listed in Table 1 is a nominal earth model, whose values are preferred for regions such as Greenland that are made up of very old cratonic units (e.g. Lambeck et al., 1998). The second source of uncertainty is from the ice models used to describe not only the GIS, but also the other major areas of glaciation. For the non-GIS, this is dealt with by using variants of the nominal ice models that contain 90% or 110% of the nominal’s ice volume (i.e. ±10%). This assumes that the uncertainties in the spatial distribution of the changes in these ice sheets are of negligible importance by distances from their centres of loading such as Greenland.

The computer programs and ice models used in this work were developed progressively at the Research School of Earth Sciences (RSES), the Australian National University, and have been applied to many GIA and sea-level change studies (e.g. Nakada and Lambeck, 1987; Lambeck et al., 1990; Johnston, 1993; Lambeck, 1995; Fleming et al., 1998; Lambeck et al., 1998; Johnston and Lambeck, 1999; Yokoyama et al., 2000). The computer programs accommodate moving coastlines, transitions between grounded and floating ice and changes in the Earth’s rotation tensor.

### Table 1

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Lower Limit</th>
<th>Upper Limit</th>
<th>Nominal</th>
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<tbody>
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<td>100</td>
</tr>
<tr>
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<td>$\eta_{\text{um}}$ ($\times 10^{20}$ Pa s)</td>
<td>2</td>
<td>5</td>
</tr>
<tr>
<td>Lower mantle viscosity</td>
<td>$\eta_{\text{lm}}$ ($\times 10^{23}$ Pa s)</td>
<td>5</td>
<td>20</td>
</tr>
</tbody>
</table>

#### 3. Sea-level change around Greenland since the LGM

The dominant contribution to sea-level change around Greenland since the LGM is glacio-isostatic emergence resulting from the Earth’s response to the retreat of the expanded GIS. The manner in which this emergence varied about the island, itself a function of the extent to which the GIS expanded, is well described by the spatial variability of the local marine limits. Marine limits are the occurrence at a given locality of the highest indication of past sea level (e.g. Funder and Hansen, 1996), as shown for Greenland in Fig. 1. Such diagrams are not descriptions of relative sea level for a given epoch, since the timing of the marine limits’ formation varies from site to site, depending upon each locality’s glacial history. There is also the problem that most marine limits lack suitable fossil material to allow the dating of their formation. Therefore, interpolations from lower, younger events are often used, with a subsequent loss of accuracy (Section 6). This point will be discussed further in Section 6. Despite these limitations, marine limits are still a useful qualitative tool in that they provide upper/lower limits when comparing predictions with observations.

The Greenland marine limits resemble elongated domes lying approximately parallel to the present-day coastline. Holocene emergence around Greenland is generally less than 100 m relative to present-day sea level, although three areas display prominent higher limits; Kong Frederik IX’s Land in West Greenland, where the limits reach up to 140 m, Scoresby Sund in East Greenland (ca 135 m) and Northwest Greenland around Hall Land (ca 120 m). A fourth area may exist in Peary Land, North Greenland, but the available data does not allow a definite statement to be made (Weidick, 1976; Funder, 1989; Funder and Hansen, 1996). These areas also correspond to the more extensive expanses of ice-free land, where the retreat of the LGM ice sheet (i.e. change in ice loading) was the greatest. For example, in Kong Frederik IX’s Land (Fig. 1) the ice-free area extends for 150–200 km between the current GIS margin and the present-day coastline. On the other hand, less extensive ice-free areas record a smaller Holocene emergence. For example, the Frederikshåb and Julianeåb districts of Southwest Greenland have marine limits between 40 and 60 m with ice-free zones of ca 50 km (Weidick, 1975; Kelly, 1977).

Fig. 1 also presents several examples of local sea-level curves from around Greenland. The dominance of the glacial-rebound contribution is apparent from the observed emergence/recession. Holocene transgressive/regressive events may have interrupted the general trend of falling sea level in some areas (e.g. Jameson Land, East Greenland, Björck et al., 1994) although evidence for such events is not definitive. Most local sea-level curves from other localities show the same general
tendencies, with coastal areas exhibiting earlier, smaller emergence than sites further inland (e.g. Weidick, 1996). Some variation in the timing and rate of emergence is apparent around the island, however the available observations generally lack the precision required to allow more detailed discussions. In most areas, sea level appears to have been close to the present-day level between ca 4–14C ka (Kelly, 1985), although in other localities this occurred much earlier (e.g. South Greenland, ca 914C ka; Bennike et al., 2002), after which it fell to below the present-day level before a late-Holocene transgression. During this period, the GIS in some areas readvanced from its Holocene minimum position, which was sometimes located behind the current ice margin. This period is termed the neoglacial and was the result of a cooling trend in the climate (e.g. Kelly, 1980). More recent evidence for sea-level fluctuations include the inundation of Eskimo and Viking sites and repeated geodetic surveys of water markers (e.g. Saxov, 1958, 1961; Weidick, 1996; Rasch and Jensen, 1997; Kuijpers et al., 1999).

Fig. 2 shows the distribution of relative sea-level data for Greenland that are considered useful for this work. Details of these data and a complete reference list is available on a website (see the Acknowledgments). There are large areas where no useful observations are available, e.g. the southeast coast adjacent the Danmark Strait and the west coast along Melville Bugt, hence few comments can be made about the glacial and sea-level histories of these areas using the procedure followed in this work. The potential for biases arising from the uneven distribution of data is also apparent, with areas such as Scoresby Sund and Disko Bugt being more densely sampled that most parts. Fig. 3 presents for each of the regional subdivisions the temporal distribution of the data. There is a relatively well-defined time from when observations are available (ca 1014C ka), marking when the present-day coastal areas became ice free. Therefore, like those areas without observations, few conclusions can be drawn about Greenland’s sea-level and glacial histories prior to this date. All together, just over 600 data have been used in this work, the vast majority of which are fossil molluscs (>80%), that mostly provide only a lower limit for sea level (Kelly, 1973; Bennike and Weidick, 2001). However, in recent years additional data from isolation basins have become available, allowing sea-level histories to be better constrained (e.g. Björck et al., 1994; Anderson and Bennike, 1997; Long et al., 1999; Bennike et al., 2002).

4. Isostatic effect of ice sheets outside of Greenland

Changes in ice loading outside of Greenland affects local relative sea level in two ways (Eq. (3)). The first is the Δsef contribution from those ice sheets. The second
is the associated isostatic response, $\Delta z_{\text{non}}$, with an uncertainty, $\delta z_{\text{non}}$, that arises from unknowns in the imperfectly defined ice and earth models. Fig. 4 presents for several sites around Greenland time-series curves of the isostatic response since the LGM arising from such ice-load changes. The curves were calculated using nominal ice models representing the North American, European and Antarctic ice sheets, the response from the sum of these ice masses, and the nominal earth model (Table 1). The contribution from the expanded Icelandic Ice Sheet has been neglected owing to its small size and because its response is restricted to the immediate vicinity of the island due to the much thinner lithosphere and underlying low-viscosity zone (e.g. Wolf et al., 1997).

The dominance of the North American isostatic contribution, owing to the great size and proximity of these ice sheets, is the most obvious result, especially when its contribution is compared with the summed non-Greenland response. There is a relatively large spatial variation in this part of the signal, for example Hall Land (Fig. 4d) experiences an isostatic response from the North American ice sheets of ca 33 m at 18 $^{14}$C ka, decreasing to <1 m at 10 $^{14}$C ka, while at Jameson Land (Fig. 4b), these values are 26 and 7 m, respectively. The European ice sheets also contribute, with noticeable spatial variability between, for example, Jameson Land and Claushavn (Fig. 4c). Another feature of these modelled curves, especially the North American and European results, is the complex temporal variation. The earlier positive values are heavily influenced by the direct gravitational attraction of the ice sheets, while the later negative values are a consequence of the collapse of the peripheral forebulges surrounding the ice sheets. Little spatial dependence is found in the Antarctic results because Greenland is in the far-field...
relative to Antarctica. Most variation in the Antarctic contribution is a function of the water load, which is in turn dependent upon each site’s location relative to the coast. The combined effect of these ice sheets is such that, in the absence of changes in the GIS, all of Greenland would today be experiencing a rise in sea level of the order of several mm yr\(^{-1}\) (e.g. Tarasov and Peltier, 2002).

The total isostatic contribution, \(\Delta z_{\text{non iso}}\), of these ice sheets for several epochs is shown in Fig. 5. The spatial and temporal variability in this response is now clear, where at 18 \(^{14}\)C ka (Fig. 5a), \(\Delta z_{\text{iso}}\) ranges between ca 0 and 25 m across Disko and Disko Bugt (west to east), and ca 45–30 m along Scoresby Sund (west to east), while reaching up to 100 m in the northwest. By 10 \(^{14}\)C ka (Fig. 5b), \(\Delta z_{\text{iso}}\) is between ca −35 and −15 m around Disko Bugt and between ca 10 and 0 m along the east coast. At 6 \(^{14}\)C ka (Fig. 5c), it is ca −40 m for Disko Bugt and −10 m for Scoresby Sund, with between ca −12 to −4 m, west to east across the island, still remaining by 2 \(^{14}\)C ka (Fig. 5d). As described in Section 2, the uncertainty associated with the non-Greenland response, \(\delta z_{\text{iso}}\), is found using a series of earth- and ice-model parameters (Table 1) with \(\delta z_{\text{iso}}\) set to equal one third of the range (1 sigma) between the maximum and minimum results.

5. Preliminary Greenland ice models

Four preliminary models of the GIS are used. These are divided between maximum and minimum versions, where maximum and minimum refers to the extent of the GIS during the LGM. For each of these, two different deglaciation starting times are considered; (i) 18 \(^{14}\)C ka, corresponding to deglaciation beginning immediately after the LGM and (ii) 14 \(^{14}\)C ka, which corresponds to the time from when deglaciation is believed to have started in West Greenland (Kelly, 1985). The minimum and maximum reconstructions of Denton and Hughes (1981) are used to define the extent of the GIS during the LGM. These are based on the flow equations of Hughes (1981), taking into account bedrock topography, isostatic displacement and different basal conditions. The minimum models employ the assumption that the Arctic ice sheets were out of phase with the more southernly ice masses and hence deprived of moisture during the maximum glaciation, while the
maximum models assume that the Arctic ice sheets were in phase with those in the south. The minimum GIS expanded to the edge of the present-day coastline during the LGM, while the maximum reached the edge of the continental shelf. For the purpose of this work, the actual mass-loss mechanisms (melting, calving) are irrelevant and only the fact that the GIS retreated and thinned is of concern.

The LGM part of these models was digitised from the ice-thickness maps presented in Hughes et al. (1981). These maps were then compared with the present-day GIS (Simon Ekholm, pers. com., National Survey and Cadastre, Denmark, Ekholm, 1996) to determine the change in ice thickness. For all models, the 10 14C ka margin is located approximately around the present-day coastline, using the results of Andersen (1981) and Funder and Hansen (1996). The ice margin is defined to have reached its current position in all areas by 7 14C ka. The ice margins between the LGM and 10 ka were linearly interpolated; 16, 14 and 12 ka for deglaciation starting at 18 14C ka, and 12 ka for the 14 14C ka starting date, with the LGM ice sheet used...
between 18 and 14 ka. Similarly, the ice margins for 9 and 8 ka were interpolated between the 10 and 7 ka margin, with some modifications from Andersen (1981). The final models after gridding have a resolution of 0.25° latitude and 0.5° longitude (ca 25 km) and are denoted as: (i) MIN1 and MIN2, which are the minimum models with 18 \(^{14}\)C and 14 \(^{14}\)C ka start-of-deglaciation times, respectively, and (ii) MAX1 and

Fig. 5. The isostatic contribution ($\Delta z_{\text{non-iso}}$, Eq. (3)) from the sum of the nominal non-Greenland ice models and the nominal earth model (Table 1) for; (a) 18 \(^{14}\)C ka, (b) 10 \(^{14}\)C ka, (c) 6 \(^{14}\)C ka and (d) 2 \(^{14}\)C ka. Contours are metres relative to the present-day sea level.
MAX2, which are the corresponding maximum models. Two important simplifications are; (i) no readvances or pauses in deglaciation are included and (ii) ice-margin retreat associated with the neoglaciaion is not considered. This deglaciation chronology is relatively simple compared to the present state of knowledge (e.g. Bennike and Björck, 2002), but it is believed to reflect those primary characteristics of the expanded GIS with which the relative sea-level signal is sensitive.

Fig. 6 shows the $\Delta_{\text{esl}}$ and change-in-ice-volume curves for each model and Fig. 7 presents the changes in ice thickness. The largest change in ice thickness occurs in those areas with the higher marine limits (compare Fig. 7 with Fig. 1). It is also seen that in some regions, the maximum model actually leads to a smaller change in ice thickness than the minimum model. This is due to ice streams being more important in the formulation of the maximum models than in the minimum, resulting in ice flowing around the coastal mountains. Another difference between the minimum and maximum models is the incorporation of the coalescence of the Innuitian Ice Sheet and the northwest GIS in the maximum models, while in the minimum models these ice sheets remain separate. While the question of whether this occurred has been the topic of much debate, the consensus today is for a model that is closer to the maximum description (e.g. England, 1999).

6. Comparisons between predictions and observations of sea-level change

Table 2 presents the statistical results found when comparing the sea-level observations with predictions (Eq. (3)), using the preliminary GIS models (Section 5) and the ice and earth models discussed in Section 4. Listed are the initial variances for the unscaled predictions ($\sigma^2_{\beta_{\text{gis}}}$, Eq. (4)), the $\beta_{\text{gis}}$ values that lead to a minimum variance between observations and predictions, and the resulting $\sigma^2$ values after scaling for each dataset subdivision (Fig. 2). For most combinations of datasets and GIS models, a reduction in the change in ice thickness as defined by the preliminary GIS models is necessary. The minimum models require a proportionally larger reduction, remembering that in some areas, the changes in ice thickness defined in the minimum models are greater than in the maximum. For all datasets, the 14 $^{14}$C ka start-of-deglaciation option results in smaller variances, while the minimum model better represents most areas. As expected from the discussion in Section 2, a single scaling parameter for the entire ice sheet is inappropriate, with different $\beta_{\text{gis}}$ values and preliminary models better suiting each area. For example, when considering the full Greenland dataset (Table 2, line 1), the single-parameter scaled maximum models better describe Greenland’s sea-level history (i.e. lower $\sigma^2$), but when the northwest dataset is excluded (Table 2, line 2), the minimum models lead to smaller variances.

These results are described further by Fig. 8, where predictions from the MIN2 (Fig. 8a) and MAX2 (Fig. 8b) models, scaled by factors derived using the regional datasets (Table 2, lines 3, 8, 14 and 17), are compared with observations. Despite the large scatter, regional differences can be identified. The most obvious are for the northwest dataset, with MAX2 giving the better fit. Fig. 9 presents the residuals between the observations and scaled predictions, again for MIN2 (Fig. 9a) and MAX2 (Fig. 9b). In general, the modelled isostatic responses are too small at earlier times (ca 10 $^{14}$C ka) and excessive at later epochs (ca 8–6 $^{14}$C ka), especially for East Greenland. This suggests that temporal, as well as spatial, scaling factors are required.

6.1. Southwest Greenland

Fig. 10 compares the observations of sea-level change and marine limits at representative sites for Southwest Greenland (Fig. 2) with the corresponding predicted curves. The predictions were found using the isostatic response from the preliminary models after applying the sector-derived $\beta_{\text{gis}}$ values (Table 2, lines 4–7). The converging of results from all models indicates that the methodology is correcting the initial predictions in a way that improves the fit with observations. However, the lack of data from earlier epochs prevents constraints to be placed on the glacial histories for those periods. For example, sites such as Middle and Inner Søndre Strømfjord (Figs. 10b and c) and Arveprinsen Ejland (Fig. 10f), where similar Holocene sea levels are predicted, display quite different LGM and early post-glacial relative sea levels (ca 18–12 $^{14}$C ka).

As discussed in Section 3, sea-level regression dominators the period covered by the available data, which in general has been reproduced by the models. The sudden change in the modelled rate of sea-level
Fig. 7. The change in ice thickness relative to the present day as defined by the (a, c) minimum (MIN1) and (b, d) maximum (MAX1) Greenland ice models at (a, b) 18 and (c, d) 10¹⁴C ka. The 14¹⁴C ka start-of-deglaciation models are not shown since the difference between the 10¹⁴C ka models associated with each starting-time option is small.
Table 2

Statistics for comparisons between observations and predictions (Eqs. (3) and (4)) of relative sea-level change arising from the four preliminary Greenland ice models. The listed datasets are subdivisions of the main Greenland dataset (Fig. 2).

<table>
<thead>
<tr>
<th>Dataset (number of observations)</th>
<th>MIN1</th>
<th>MAX1</th>
<th>MIN2</th>
<th>MAX2</th>
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<tbody>
<tr>
<td></td>
<td>$\beta_{\text{gis}}$</td>
<td>$\sigma^2$</td>
<td>$\beta_{\text{gis}}$</td>
<td>$\sigma^2$</td>
</tr>
<tr>
<td>----------------------------------</td>
<td>--------</td>
<td>--------</td>
<td>--------</td>
<td>--------</td>
</tr>
<tr>
<td>1 Full (625)</td>
<td>14.8</td>
<td>0.90</td>
<td>14.3</td>
<td>12.7</td>
</tr>
<tr>
<td>2 Excluding Northwest (536)</td>
<td>8.7</td>
<td>0.85</td>
<td>7.5</td>
<td>9.5</td>
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<tr>
<td>3 Southwest (221)</td>
<td>10.2</td>
<td>0.92</td>
<td>9.9</td>
<td>12.1</td>
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<tr>
<td>4 South (47)</td>
<td>31.3</td>
<td>1.19</td>
<td>30.7</td>
<td>31.1</td>
</tr>
<tr>
<td>5 Central south (58)</td>
<td>4.6</td>
<td>0.91</td>
<td>4.2</td>
<td>10.6</td>
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<tr>
<td>6 Central north (50)</td>
<td>5.4</td>
<td>0.81</td>
<td>3.1</td>
<td>5.6</td>
</tr>
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<td>7 North (66)</td>
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<td>5.4</td>
</tr>
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<td>8 East (161)</td>
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<td>7.4</td>
<td>6.9</td>
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<tr>
<td>9 South (25)</td>
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<td>5.5</td>
<td>10.8</td>
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<td>0.72</td>
<td>4.1</td>
<td>7.8</td>
</tr>
<tr>
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<td>7.6</td>
<td>9.9</td>
</tr>
<tr>
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<td>0.84</td>
<td>2.8</td>
<td>3.6</td>
</tr>
<tr>
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<td>11.2</td>
<td>1.44</td>
<td>5.2</td>
<td>3.8</td>
</tr>
<tr>
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<td>0.82</td>
<td>3.7</td>
<td>8.7</td>
</tr>
<tr>
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<td>2.6</td>
<td>10.1</td>
</tr>
<tr>
<td>16 South (89)</td>
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<td>0.85</td>
<td>4.6</td>
<td>7.7</td>
</tr>
<tr>
<td>17 Northwest (89)</td>
<td>52.1</td>
<td>1.89</td>
<td>40.0</td>
<td>32.4</td>
</tr>
<tr>
<td>18 South (43)</td>
<td>67.0</td>
<td>0.00</td>
<td>0.0</td>
<td>35.3</td>
</tr>
<tr>
<td>19 North (46)</td>
<td>39.3</td>
<td>1.90</td>
<td>15.4</td>
<td>30.4</td>
</tr>
</tbody>
</table>

Fig. 8. Comparisons of observed and scaled-predicted relative sea-level change resulting from the; (a) MIN2 and (b) MAX2 preliminary Greenland models. The predictions are scaled using $\beta_{\text{gis}}$ (Eq. (3)) derived from the regional datasets (Table 2, lines 3, 8, 14 and 17). The straight line represents the 1:1 relationship between observations and predictions.
change at 714°C ka is due to the deglaciation of the GIS being completed at that time, as defined by the preliminary models. If a neoglacial component is considered, then the curves from 7 14°C ka onward would be displaced downwards, hence providing a better fit at several sites for mid- to late-Holocene times, e.g. Inner Sondre Strømfjord and Arveprinsen Ejland.

The change in ice thickness around the southern-most site, Kap Farvel (Fig. 10a) is clearly underestimated, with the predicted curves being much lower than the observations. Because the continental shelf in this area is fairly close to the present-day coastline (ca 50 km), the models' profiles need to incorporate a greater change in ice thickness towards the seaward end of the ice. This area is also believed to have deglaciated at a faster rate than has been defined in the preliminary models (Bennike et al., 2002). For Middle Sondre Strømfjord, an increase in the change in ice thickness is required since the curves do not reach the two higher observations and barely meet the marine limit. However, this area is prone to the upward displacement of marine material by glacial action (Sugden, 1972), and such interpretations need to be treated with caution. At Inner Sondre Strømfjord, the modelled curves are higher than the observations during the mid-Holocene, consistent with the observations being lower limits. However, as discussed above, the retreat of the GIS to behind its current margin would improve these results. Holsteinborg (Fig. 10d) also suggest an inadequate change in local ice loading after scaling, a result that is repeated at West Central Disko (Fig. 10e). The ice sheet's profile in these areas either needs to be modified as for Kap Farvel, or/and, incorporate more ice some distance offshore, but not as far as the shelf edge. Arveprinsen Ejland shows a similar pattern as Inner Sondre Strømfjord. The intersection of the recessional part of the modelled curves with the marine limit at Inner Sondre Strømfjord and Arveprinsen Ejland provides estimates for the age of these features, although it is still possible that they formed during an earlier period of relatively stable sea level.

6.2. East Greenland

Fig. 11 compares scaled predictions and observations, including marine limits, from selected East Greenland sites (Fig. 2, Table 2, lines 9 to 13). The previously discussed temporal dependence observed in the residuals (Fig. 9) becomes more obvious. For example, at Mesters Vig (Fig. 11b), the predictions at ca 7 14°C ka are above the observations, but the earlier epoch results, ca 9 14°C ka, lie below them. Following the same trend as Southwest Greenland, a greater change in ice thickness along coastal areas and/or extra ice offshore is required (e.g. Mesters Vig) while sites closer to the present-day margin (e.g. Waltershausen Gletscher, Fig. 11c) appear adequately described.

As mentioned in Section 3, some areas show evidence for transgressive/regressive events during the Holocene. While glacial readvancements are often proposed to explain such events, another possibility is that they are a result of the relative importance of the isostatic response due to both local and non-Greenland load changes and the equivalent sea-level rise. Such events are seen to be theoretically possible using these models (e.g. Jameson Land, Fig. 11a).

The northern sector (e.g. Hvalrosodden, Fig. 11d) displays a different behaviour to much of East Greenland, with the maximum models being preferred (Table 2, line 13, Fig. 11d). It has been proposed that this area experienced a later deglaciation than other parts of Greenland, with the GIS expanding some distance offshore, since the shelf in this area extends ca 300 km (Bennike and Weidick, 2001). There is also a recently discovered island (Tobias Øer) that may have formed a base for additional ice buildup (Bennike and Björck, 2002). Our results add support to that hypothesis, also considering that in the northern part of this sector is the
Fig. 10. Comparisons of observations and scaled-predicted relative sea-level curves (Table 2, lines 4–7) for various sites around Southwest Greenland (Fig. 2). Dark-grey shading indicates the estimated local marine limits (Fig. 1). Symbols: triangles—lower limits, diamonds—mean sea level, inverted triangles—upper limits.
outlet for the Northeast Ice Stream (Fahnestock et al., 1993; Thomsen et al., 1997), which exhibits frequent fluctuations in the extent of the surrounding outlet glaciers (Higgins, 1991).

6.3. Northeast Greenland

Fig. 12 presents the predictions, observations and marine limits for Northeast Greenland (Fig. 2, Table 2, lines 15 and 16). Jørgen Brønlund Fjord and Kap København (Figs. 12b and c) are best represented by these predictions, while the results are less favourable for Peary Land and Holm Land (Figs. 12a and e). As with the previous regions’ results, the models include a deficiency in the change in ice volumes required along the present-day coast (e.g. Peary Land and Holm Land). Predictions for areas closer to the present-day margin have either an excessive change in ice loading (e.g. Danmark Fjord and Blåso, Figs. 12d and f) or are adequately described (e.g. Jørgen Brønlund Fjord).

Overall, the minimum ice models appear to best serve this region. This is consistent with work that describes how the LGM margin was close to the present-day coast along Holm Land (Hjort, 1997), although as discussed previously, the LGM margin to the south was possibly located further offshore (north East Greenland, Fig. 11d). On the other hand, ice shelves are believed to have existed off East Peary Land and Johannes V. Jensen Land in the Wandal Sea, suggesting a greater ice loading, which would be better for sites such as Peary Land (Funder and Hjort, 1980).

6.4. Northwest Greenland

Modelled sea-level curves from selected Northwest Greenland sites (Fig. 2) are compared with observations and local marine limits in Fig. 13 (Table 2, lines 18
Fig. 12. Comparisons of observations and scaled-predicted relative sea-level curves (Table 2, lines 15 and 16) for various sites around Northeast Greenland (Fig. 2). Dark-grey shading indicates the estimated local marine limits (Fig. 1). Symbols: triangles—lower limits, diamonds—mean sea level, inverted triangles—upper limits.
and 19). The minimum-model results for Thule (Fig. 13a) and Iterluk (Fig. 13b) are completely inadequate and are not included since the resulting scaling factors are meaningless (Table 2, line 18). This supports the current view that the Innuitian and Northwest Greenland ice masses coalesced during the LGM (e.g. England, 1998, 1999; Zreda et al., 1999). However, the fits between the observations and maximum model results are also poor, and require significant modifications to the spatial and temporal characteristics of the models. In these later models, the present-day coastline was cleared of ice by 10 $^{14}$C ka, whereas the more likely deglaciation chronology involved a largely latitudinal movement of the ice margin through either end of Nares Strait (Fig. 1). This started ca 10.1 $^{14}$C ka in the north and 9 $^{14}$C ka in the south, producing a clear ocean way by 7.5 $^{14}$C ka (England, 1999).

Regarding the northern-sector sites, Hall Land (Fig. 13c) and Nyeboe Land (Fig. 13d), the scaled maximum models produce reasonable fits with the observations, although still underestimating the response required at ca 8 $^{14}$C ka. It must also be remembered that the non-GIS, especially the Innuitian Ice Sheet, have a significant influence in this area (see Fig. 6d). Hence, future GIS models will require the local glacial history to be examined in conjunction with that of the Canadian Arctic (e.g. Dyke, 1999).

7. The first-order Greenland ice model—GREEN1

A first-order model for the GIS has been produced in an interactive process, based on the analysis described in Section 6. The first step involves defining for each sector the preliminary model that results in the lowest $\sigma^2$ after scaling (Table 2). We find that the 14 $^{14}$C ka start-of-deglaciation results in the smallest variances, with the scaled MIN2 model best reproducing the sea-level
history for most areas. The exceptions are Northwest Greenland, southern-most Southwest Greenland and parts of East Greenland, all of which are better represented by the scaled MAX2 model. The corresponding $\beta_{\text{GIS}}$ values are applied to the appropriate preliminary model, and these new sector models are combined to produce an “intermediate” ice model. Predictions from the intermediate model are then compared with observations, with the resulting $\beta_{\text{GIS}}$ values applied to each sector of the intermediate model to produce the first-order model, GREEN1 (Table 3).

The appropriate model for the centre of the GIS and coastal areas without observations is found by comparing the observations with predictions based on iterations of GREEN1 incorporating MIN2 and MAX2 portions of those areas, scaled by 1.0, 0.9, etc. Based on these comparisions, we select the MIN2 model’s portion, scaled by 0.7 although we noted little difference for values of 0.8–0.6. The final $\beta_{\text{GIS}}$ values that result from the use of GREEN1 tend to be around unity (Table 3), indicating: (i) if another iteration were carried out, the improvement would be minimal and, (ii) this model provides the optimum agreement between predictions and observations, given the available datasets and assumptions made about the Earth’s rheology.

Fig. 14 shows the GREEN1 model for; (a) the LGM (18–14 $^14$C ka) and (b) 10 $^14$C ka, and the resulting topography after considering sea-level change for; (c) the LGM (18 $^14$C ka) and (d) 10 $^14$C ka. The change in ice thickness since the LGM along the present-day coastline in most areas is of the order of 500 m, with changes of over 1500 m in some parts adjacent to the present-day margin. From 10 $^14$C ka, changes adjacent to the present-day margin are still $>$ 1000 m in areas with highly raised shorelines. The GREEN1 ice-volume changes are equivalent to a eustatic sea-level contribution of 3 m at the LGM and 1 m at 10 $^14$C ka, within the range predicted by recent thermomechanical ice-sheet modelling (Huybrechts, 2002).

### Table 3

The preliminary ice models and scaling factors that make up the intermediate and first-order (GREEN1) Greenland ice models. The listed datasets are subdivisions of the main Greenland dataset (Fig. 2).

<table>
<thead>
<tr>
<th>Datasets</th>
<th>Ice model</th>
<th>Scale</th>
<th>Intermediate</th>
<th>GREEN1</th>
</tr>
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</table>
|            | $\beta_{\text{GIS}}$ | $\sigma^2$ | $\beta_{\text{GIS}}$ | $\sigma^2$
<table>
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<tr>
<td>Full Dataset</td>
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<td>5.5 1.00</td>
<td>5.4 1.00</td>
<td>5.4 1.00</td>
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<td></td>
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<td></td>
</tr>
<tr>
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<td>MAX2 1.70</td>
<td>6.7 1.02</td>
<td>15.3 1.11</td>
<td>14.9 1.04</td>
</tr>
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<td>13.8 1.05</td>
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<tr>
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<td>2.3 0.92</td>
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<tr>
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<td>14.9 1.04</td>
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<td>East</td>
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<td></td>
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<tr>
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<td>6.4 1.04</td>
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<tr>
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8. Discussion

Fig. 15 presents the modelled $\Delta z_{\text{obs}}$ (Eq. (1)) over Greenland resulting from GREEN1 and the other model parameters used in the previous sections. We see that the general spatial form of the areas of maximum sea-level recession correspond very well to the higher marine limits (Fig. 1). We however note discrepancies in the modelled elevations. For example, in the southeast and along Melville Bugt, no Holocene recession is reproduced whereas marine limits with elevations up to 20 m are found, although the marine-limit estimates in that area may be unreliable (Bennike, pers. com.). Likewise, the highest predicted sea levels in areas such as Disko Bugt and Scoresby Sund are lower than the observed marine limits.
Fig. 14. (a, b) The first-order Greenland ice model, GREEN1, showing the change in ice thickness since; (a) 18 $^{14}$C ka and (b) 10 $^{14}$C ka. (c, d) The topography of Greenland, accommodating the changes in ice thickness and sea level at (c) 18 $^{14}$C ka and (d) 10 $^{14}$C ka.
In Fig. 16, the range in $D_{z^{\text{rsl}}}$ is presented for representative sites selected from Figs. 10–13. Large uncertainties occur for earlier times (between ca 18 and 9 $^{14}$C ka), of the order of ±20 m, emphasizing the low resolution for the early epochs where the sea-level data is of an uncertain quality. Also of note is that the
Fig. 16. Comparisons of observations and the range in predicted relative sea level using the GREEN1 model and a series of non-Greenland ice and earth models (Table 1) for various locations around Greenland. The ±1σ uncertainty (light-grey shading) is plotted over the nominal results. Dark-grey shading indicates the estimated local marine limit (Fig. 1). Symbols: triangles—lower limits, diamonds—mean sea level, inverted triangles—upper limits.
observed sea levels below the present-day level are not reproduced by the model. To investigate this later point, a series of order-of-magnitude calculations have been made using an axisymmetric GIA model (e.g. Johnston, 1993; Bennike et al., 2002) in which the present-day ice model has the same area and equivalent-water content as the current GIS. The expanded ice sheet is constrained by the GREEN1 $\Delta z_{\text{gis}}$ values, with the ice-sheet profiles being simple parabolas. A standard model that excludes a neoglacial component is compared with three other models; (i) NEO1, where the GIS retreats to behind the present-day margin between 7 and 6 $^{14}$C ka by 10 km, readvancing to the present-day margin between 3 $^{14}$C ka and today, (ii) NEO2, which has the same timing but a 20 km movement and (iii) NEO3, which has a 20 km movement but readvances from 1 $^{14}$C ka.

Fig. 17 presents the results of these tests for; (a) the axisymmetric ice models at different epochs and (b) the associated change in ice loading relative to the present day. Figs. 17c and d shows the predicted $\Delta z_{\text{gis}}$ curves (Eq. (3)) arising from these ice models using the nominal earth model (Table 1) at two locations; (c) the present-day ice margin and (d) the present-day coastline. The neoglacial contribution is shown in Figs. 17e and f where (e) is the present-day ice margin and (f) the present-day coastline. The downward displacement of sea level at late-Holocene times by the neoglacial contribution demonstrates that such ice-sheet behaviour is a possible mechanism for the observed fall in sea level to below the present-day level and subsequent transgression.

These neoglacial results are now combined with the GREEN1 model for the case of Arveprinsen Ejland, West Greenland, a site whose sea-level history is reasonably well described by GREEN1 (Figs. 10f and 17b). Long et al. (1999) comment that sea level at 2.5 $^{14}$C ka in this area was ca $\sim$ 5 m, while the predicted value from GREEN1 is ca 6 m. Since the NEO2 model predicted a downward shift in sea level at this time of ca 5 m, results from an axisymmetric ice model that incorporates an ice-margin movement of 40 km with the readvance starting at 2.5 $^{14}$C ka are combined with those from the GREEN1 model. These results are shown in Fig. 18; (a) the predictions from the GREEN1 model, (b) the effect of the new neoglacial description with respect to the standard model and (c) the sum of the initial predictions and neoglacial response. The fall in sea level to below today’s level and the subsequent transgression are reproduced, but the early-Holocene predictions are now less representative of the observations. Hence, the introduction of a neoglacial component in the GREEN1 model impacts on the solution for the ice model for earlier epochs. Nonetheless, the estimated neoglacial ice-margin movement is consistent with studies based on moraine dating (10’s of km of ice margin movement, vanTatenhove et al., 1996) and vertical motion observed by GPS measurements (ca 50 km, Wahr et al., 2001a, b).

9. Conclusions

This paper deals with the development of a first-order model of the ice-load history of the GIS since the LGM. The first part is an assessment of the contribution to Greenland’s sea-level history of changes in the ice sheets located outside of Greenland. The findings of this section are:

- Changes in these ice sheets contributed a complex spatial and temporal signal to Greenland’s sea-level history, of the order of metres to 10’s of metres (Figs. 4 and 5).
- The North American ice sheets are the dominant part of this signal, especially the Innuitian Ice Sheet’s influence on Northwest Greenland.
- The ongoing isostatic contribution of these ice-load changes, if treated in isolation, causes an ongoing rise in sea level around Greenland today of the order of several mm yr$^{-1}$.

Inversions of sea-level change along the Greenland coast are made using different initial GIS ice models (Figs. 6 and 7) and scaling parameters ($\beta_{\text{gis}}$, Eq. (4)) that minimise the variance ($\sigma^2$, Eq. (5)) between predictions and observations of relative sea level. We find that:

- A single scaling parameter for all of Greenland is inappropriate, hence requiring the spatial subdivision of the observational dataset (Table 2).
- A deglaciation that started at 14 $^{14}$C ka provides better results (smaller $\sigma^2$) than an earlier (18 $^{14}$C ka) date.
- The scaled minimum (MIN2) models better reproduced Greenland’s sea-level history for most areas. The exceptions are southern-most Southwest Greenland, parts of East Greenland and Northwest Greenland. The coalescence of the Greenland and Innuitian ice sheets is reaffirmed.
- The LGM ice sheet extended some distance offshore, but not usually as far as the shelf edge. However, this probably did occur in some areas, namely southern-most West Greenland, northern East Greenland and Northwest Greenland.

The resulting first-order Greenland ice model, GREEN1, has the following characteristics:

- A deglaciation starting at 14 $^{14}$C, although a single starting date for the entire island would be too simplistic.
- LGM and 10 $^{14}$C ka equivalent sea-level contributions of 3.1 and 1.9 m, respectively.
Fig. 17. Effect of a neoglacial contribution on glacial-isostatic rebound in Greenland. See the text for model descriptions. (a, b) Profiles at various epochs of: (a) ice-sheet heights and (b) changes in ice thickness. (c, d) Isostatic response for sites located at: (c) the present-day ice margin and (d) the present-day coastline. (e, f) Differences between results with and without neoglacialization at: (e) the present-day ice margin and (f) the present-day coastline.
Changes in ice thickness of the order of 500 m along the present-day coast and > 1500 m adjacent to the current margin in areas of major ice retreat. The resulting sea-level histories from the use of this model are characterised by:

- The spatial form of the marine limits have been reproduced (compare Figs. 1 and 15). However, the observed limit elevations are not always reached, although there are doubts as to the reliability of some marine limits.
- Mid- to late-Holocene sea levels below today’s level are not reproduced (Fig. 16).

Regarding the last point, to reproduce sea levels that are below the present day’s and the subsequent late-Holocene transgression, a neoglacial component where the GIS retreated ca 40 km behind its present-day margin and readvanced over the last few thousand years is required (Fig. 18). Such a contribution, if included in an ice model, would also affect predictions for earlier times.

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