Late Weichselian relative sea-level changes and ice sheet history in southeast Greenland

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ABSTRACT

Relative sea-level (RSL) observations from the margins of the Greenland Ice Sheet (GIS) provide information regarding the timing and rate of deglaciation and constraints on geophysical models of ice sheet evolution. In this paper we present the first RSL record for the southeast sector of the GIS based on field observations completed close to Ammassalik. The local marine limit is c. 69 m above sea-level (asl) and is dated to c. 11 k cal. yrs BP (thousand calibrated years before present) and is a minimum date for ice free conditions at the study site. RSL fell to c. 24 m asl by 9.5 k cal. yrs BP and continued to fall at a decreasing rate to reach close to present by 6.5 k cal. yrs BP. Our chronology agrees with radiocarbon dates from offshore cores that indicate ice free conditions on the adjacent mid-shelf by 15 k cal. yrs BP. We compare the new RSL data with predictions generated using two recently published glaciological models of the GIS that differ in the amount and timing of ice loading and unloading over our study area. These two GIS models are coupled to the same Earth viscosity model and background (global) ice model to aid in the data-model comparison. Neither model provides a close fit to the RSL observations. Based on a preliminary sensitivity study using a suite of Earth viscosity models, we conclude that the poor data-model fit is most likely due to an underestimate of the local ice unloading. An improved fit could be achieved by delaying the retreat of a thicker ice sheet across the continental shelf. A thick ice sheet extending well onto the continental shelf is in agreement with other recent observations elsewhere in east and south Greenland.

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1. Introduction

Relative sea-level (RSL) data provide the most powerful available constraints on the Holocene dimensions of the Greenland Ice Sheet (GIS). Thus, the age and elevation of the marine limit (the highest level attained by RSL since deglaciation), provide information on the timing of ice margin retreat and net uplift since that time (Funder and Hansen, 1996). The rate of early and mid Holocene RSL fall provides information regarding the timing and magnitude of ice unloading, whilst a late Holocene rise in RSL observed in west and south Greenland reflects the combined effects a readvance of the GIS during the neoglacial, the collapse of the Laurentide Ice Sheet forebulge and ice-equivalent ‘eustatic’ sea-level change (Kelly, 1980). RSL data also constrain geophysical models of ice sheet history during the Late Weichselian that provide insights into the longer-term history of the ice sheet, including its response to climate and its contribution to global sea-level change (Bennike et al., 2002; Tarasov and Peltier, 2002; Huybrechts, 2002; Fleming and Lambeck, 2004).

Relative sea-level observations in west and south Greenland are reasonably abundant, with data points derived from radiocarbon-dated marine molluscs, drift wood or whale bones, originally deposited in beaches or glaciomarine deposits and now uplifted above present sea-level. Archaeological observations yield additional RSL estimates since c. 4 k cal. yrs BP (thousand calibrated years before present) (Rasch and Jensen, 1997; Rasch, 2000). Each of these types of data has age and altitude uncertainties that are typically ±5 to 10 m and ±200 to 400 cal. yrs. Less common are data obtained from isolation basins, natural rock basins that at various times in their history are either connected to or isolated from the sea (Foged, 1973; Bennike, 1995; Long et al., 1999, 2003, 2006; Long and Roberts, 2003; Sparrenbom et al., 2006a,b). These have better resolved age and height relationships to former sea-level, typically ±0.5 m and ±100 cal. yrs. Where these data exist from a relatively small geographical area, they provide particularly useful constraints on ice sheet history and geophysical models of glacial isostatic adjustment (GIA), which include both an Earth and ice component.

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Fig. 1. Location map of the study area showing the location of the lakes sampled, and the survey point for observations relating to the local marine limit (ML).
However, some parts of Greenland lack any RSL observations. In these areas the chronology of Holocene ice sheet dynamics is at best rudimentary and constraints on geophysical models are lacking. Notable is the stretch of coast in southeast Greenland, between Scoresby Sund and Kap Farvel (Fig. 1) which has not previously been subject to RSL study. Some offshore data exist from this sector regarding the timing of initial ice margin retreat (see below), but there is no terrestrial evidence (Bennike and Björck, 2002). We therefore lack a basic timing of initial ice margin retreat (see below), but there is no terrestrial evidence. We therefore lack a basic timing of initial ice margin retreat (see below), but there is no terrestrial evidence.

In this paper we address this deficiency by developing the first RSL curve from the southeast sector of the ice sheet. We use isolation and lake basin data collected from a site close to the town of Ammassalik (Fig.1). Our objective is to determine the approximate age of the marine limit, to define the general form of the RSL history and to compare these data with predictions based on two recently published ice models that differ in the amount and timing of ice loading and unloading over our study area. By so doing, we seek to improve our understanding of the Late Weichselian history of the GIS in this hitherto little studied region.

2. Previous work

Elevation contours of the marine limit in east Greenland define an elongate dome centred on Scoresby Sund that attains a maximum altitude of 140 m asl. The limit falls in elevation to c. 40 m asl to the south of this area but observations are limited (Funder and Hansen, 1996). The age of the marine limit in the Ammassalik area is not known, although radiocarbon dates obtained from sea bed cores from the Kangerlussuaq trough, to the north of Ammassalik, suggest that the GIS here extended to the shelf edge up until at least 17 k cal yrs BP (Miernet et al., 1992), with the middle part of the trow ice free by 15 k cal yrs BP (Jennings et al., 2002, 2006).

There are no RSL observations between the Scoresby Sund area (Björck et al., 1994; Funder and Hansen, 1996) and the southern tip of Greenland (Fredskild, 1973; Bennike et al., 2002). In the latter area, two detailed RSL studies at Nanortalik and Qaqatoq use the isolation basin methodology, with data collected from basins above and below present sea-level (Sparrenbom et al., 2006a,b). Ice-free conditions at Nanortalik, on the outer coast, are dated to c. 14 k cal yrs BP, with RSL falling from a marine limit at c. 40 m asl to reach present by c. 9.3 k cal yrs BP. RSL here has risen by c. 10 m since sometime before 5 k cal yrs BP (Sparrenbom et al., 2006a).

3. The study area

The study area comprises a fjord landscape and numerous small islands, with the coastal mountains reaching c. 1000 m asl. It is a low arctic maritime environment with a mean annual temperature of 1 °C and annual precipitation of c. 825 mm. The bedrock geology is dominated by granite–gneiss. Nearshore waters are typically ice covered from December to May. The field site is 20 km northeast of Ammassalik and was selected on the basis of the low-lying lakes that exist in this area (Fig. 1). The landscape is sparsely covered in surficial sediment and the abundant exposed bedrock has been glacially scoured by ice flowing in a general NW to SE direction. The maximum tidal range at Ammassalik is 3.66 m and highest astronomical tide (HAT) and mean high water of spring tides (MHWST) are 2.10 m and 1.64 m above mean sea-level (MSL) respectively.

4. Methods

We surveyed every lake below 150 m asl on the island of Qernertivartivit and on the adjacent mainland to its north to identify suitable coring targets, based on their size, water depth and sediment infill. Isolation basins suitable for sampling occur on Qernertivartivit up to an elevation of c. 30 m asl, but the lakes between this altitude and the local marine limit (c. 69 m asl, see below) are too shallow to contain sediments suitable for palaeoenvironmental reconstruction. Two small lakes just above the marine limit were sampled to determine a minimum age for the establishment of ice free conditions.

Lake sediments were collected using a Russian-type sampler or modified piston corer from a small tethered boat. We surveyed all elevations by closed levelling transects with a Sokisha level to a local benchmark, which was then surveyed to the local monthly high tide observed during our fieldwork and corrected to mean sea-level (MSL) by comparison with tidal data from Ammassalik provided by The Royal Danish Administration of Navigation and Hydrography. We assume a ±0.2 m height uncertainty to account for possible differences in tidal range between the study site and Ammassalik and for uncertainties in tidal predictions. We surveyed sill altitudes using a level and staff, recording the minimum and maximum elevations of each basin sill. The sill elevations are corrected to MSL based on the halobian classificaion scheme. Core chronologies are provided by Accelerator Mass Spectrometry (AMS) radiocarbon dates on thin (0.5 to 1 cm) bulk sediment slices of lake gyrtja (Table 1). The calibration programme used is CALIB 5.0.1 (Reimer et al., 2004) and all calibrated dates are cited with a two sigma age range or as the median of this range in thousands of calibrated years before present (k cal yrs BP).

5. Results

5.1. The marine limit

The marine limit is defined by the lower limit of perched boulders above wave-washed bedrock. This limit is clearly defined in the field.

### Table 1

<table>
<thead>
<tr>
<th>Basin code</th>
<th>Laboratory code</th>
<th>¹⁴C age ±1σ</th>
<th>Cal. yrs BP ±2σ</th>
<th>Max. sill altitude (m above MSL)</th>
<th>Min. sill altitude (m above MSL)</th>
<th>Reference water level</th>
<th>Indicative meaning (m)</th>
<th>MSL (m)</th>
<th>Type of date</th>
</tr>
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<tr>
<td>QT97</td>
<td>SUERC-9428</td>
<td>9659±67</td>
<td>11205–10775</td>
<td>97</td>
<td>Not relevant</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>Not relevant</td>
</tr>
<tr>
<td>QT93</td>
<td>SUERC-9426</td>
<td>9376±65</td>
<td>10766–10305</td>
<td>93</td>
<td>Not relevant</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>Not relevant</td>
</tr>
<tr>
<td>QT23</td>
<td>SUERC-9425</td>
<td>8523±67</td>
<td>9654–9411</td>
<td>23.61</td>
<td>22.40</td>
<td>MHWST-HAT</td>
<td>1.87±0.64</td>
<td>21.13</td>
<td>AMS</td>
</tr>
<tr>
<td>QT11</td>
<td>SUERC-9441</td>
<td>7594±53</td>
<td>8538–8324</td>
<td>12.13</td>
<td>11.72</td>
<td>MHWST-HAT</td>
<td>1.87±0.64</td>
<td>9.85</td>
<td>AMS</td>
</tr>
<tr>
<td>QT7</td>
<td>SUERC-9421</td>
<td>7202±51</td>
<td>8180–7942</td>
<td>7.69</td>
<td>7.35</td>
<td>MHWST-HAT</td>
<td>1.87±0.34</td>
<td>5.65</td>
<td>AMS</td>
</tr>
<tr>
<td>QT3</td>
<td>SUERC-9420</td>
<td>5868±46</td>
<td>6791–6558</td>
<td>3.33</td>
<td>2.91</td>
<td>MHWST-HAT</td>
<td>1.87±0.30</td>
<td>1.25</td>
<td>AMS</td>
</tr>
</tbody>
</table>

MSL = mean sea-level. The indicative meaning describes the height uncertainty associated with each index point.
Fig. 2. The stratigraphy of the lake basins sampled.
and ten measurements on the lower limit of perched boulders yielded an altitude of c. 69 m asl (68.75±0.82 m).

5.2. Lakes above the marine limit

5.2.1. QT97 (c. 97 m asl)
A transect of seven cores across this shallow (<2 m water depth) lake demonstrated dense blue–grey silt sands overlain by an iron-stained, silt-rich gyttja (Fig. 2). A sample from the base of the gyttja in core 5 yielded an AMS radiocarbon date of 9659±67 BP (11205–10775 cal. yrs BP, SUERC-9428).

5.2.2. QT93 (c. 93 m asl)
A single core from the centre of this lake sampled a compact grey sand silt that is overlain by a dark grey organic silt and then a dark brown gyttja containing occasional roots (Fig. 2). A sample from the base of the grey organic silt (between 285 and 284 cm) yielded an AMS radiocarbon date of 9376±65 BP (10766–10305 cal. yrs BP, SUERC-9426).

5.3. Lakes below the marine limit

5.3.1. QT23 (c. 23 m asl)
Seven cores reveal a consistent stratigraphy that comprises a grey silt sand overlain by a thin (c. 5 cm) laminated organic silt and then a light brown gyttja that extends to lake sediment surface (Figs. 2 and 3A). Diatoms from the sample core record an up-core transition from poly- to meso- and then oligohalobous taxa. The isolation contact at 744 cm is AMS radiocarbon dated to 8523±67 BP (9654–9411 cal. yrs BP, SUERC-9425).

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**Fig. 3.** Summary diatom diagrams from each of the sample cores. Diatom frequencies are classified according to the halobian classification scheme. A) QT23, B) QT11, C) QT7, D) QT3.
5.3.2. QT11 (c. 11 m asl)

The lake stratigraphy comprises a lower sand–silt that is overlain by a red/brown, faintly laminated silt gyttja (Figs. 2 and 3B). This laminated unit passes upwards into a brown, elastic gyttja that extends to the lake bed. The lithostratigraphy of the sample core shows an abrupt lithostratigraphic contact between a grey silt gyttja and the overlying soft brown gyttja at 300 cm. The diatom evidence records a mixed sample at 300 cm that contains mostly meso- and oligohalobous taxa. Above 300 cm the assemblage switches to an oligohalobous-dominated assemblage. The abrupt change in stratigraphy could indicate a hiatus at 300 cm and therefore we collected a sample from the top of the brackish water assemblage at 301–300 cm for AMS radiocarbon dating. This provides a minimum age for isolation of 7202±51 BP (8160–7942 cal. yrs BP, SUERC-9421).

5.3.3. QT7 (c. 7 m asl)

The deepest sediments in this lake are a dark brown organic-rich silt sand with remains of the common mussel (*Mytilus edulis*) (Figs. 2 and 3C). This unit contains a distinct 2–5 cm-thick, coarse sand horizon in several cores. Overlying this is a grey silt gyttja that passes upwards into a brown, elastic gyttja that extends to lake bed. The lithostratigraphy of the sample core shows an abrupt lithostratigraphic contact between a grey silt gyttja and the overlying soft brown gyttja at 300 cm. The diatom evidence records a mixed sample at 300 cm that contains mostly meso- and oligohalobous taxa. Above 300 cm the assemblage switches to an oligohalobous-dominated assemblage. The abrupt change in stratigraphy could indicate a hiatus at 300 cm and therefore we collected a sample from the top of the brackish water assemblage at 301–300 cm for AMS radiocarbon dating. This provides a minimum age for isolation of 7202±51 BP (8160–7942 cal. yrs BP, SUERC-9421).

5.3.4. QT3 (c. 3 m asl)

We collected a single piston core from the centre of this deep, steep-sided lake (Fig. 2). The sample core contains a sand silt (with unidentified fish bones and some humified organic material) which passes up into a black, finely laminated organic gyttja and then a light brown silt gyttja that extends to lake bed. Diatoms record basin isolation by an up-core replacement of polyhalobous by meso- and then oligohalobous taxa (Fig. 3D). The isolation contact is AMS radiocarbon dated to 5868±46 BP (6791–6558 cal. yrs BP, SUERC-9420).

6. Relative sea-level changes

The data described above enable us to reconstruct a RSL curve for the study area (Fig. 4). The dates from two lakes above the local marine limit provide minimum ages for the age of the local marine limit and the development of ice free conditions. The older of the two dates from QT97 suggests ice free conditions were established by c. 11 k cal. yrs BP.

We cannot define the detailed pattern of RSL change during c. 11 ka and 9.5 k cal. yrs BP due to the absence of isolation basins between the marine limit and QT23. RSL studies elsewhere in west Greenland (e.g. Disko Bugt (Long et al., 2003), Søndre Strømfjord (Ten Brink, 1974), as...
well as in Scoresby Sund (Funder and Hansen, 1996) demonstrate that RSL fell swiftly from the marine limit towards present sea-level during the early Holocene in response to rapid glacio-isostatic rebound following ice removal. We note, however, that a rapid fall in RSL following deglaciation is not always recorded in Greenland. In the Nanortalik area (Kap Farvel), for example, RSL fell slowly between c. 14 k and 12 k cal. yrs BP, before dropping more quickly during the early Holocene (Sparrenbom et al., 2006a). Moreover, on Disko Island, Ingólfsson et al. (1990) argue that RSL rose to the local marine limit before falling during the early Holocene.

Our data indicate that mean sea-level fell from 21 m to c. 1 m asl between c. 9.5 and 6.7 k cal. yrs BP. There was a pronounced slowdown in the rate of RSL fall during this period. We observe that a thin coarse sand unit recorded in the base of QT7 dates from sometime before c. 8 ka cal. yrs BP (Fig. 2). We do not know the origin of this unit, but note that the Storegga landslide and tsunami dates from c. 8.2 ka cal. yrs BP and that other authors have recently interpreted coarse, fjord-bed sediments of this approximate age in east Greenland to this source (Wagner et al., 2006). We have no RSL data after 6.7 ka cal. yrs BP to further constrain the mid and late Holocene RSL record from the study area. However, the absence of any marine incursions in the lowest lying lake basin (QT3) indicates that RSL was at or below this level for the remaining 6 k yrs of the Holocene.

7. Geophysical modelling

7.1. The models

We now compare our new RSL data to predictions computed from a glacio-isostatic adjustment (GIA) model. We consider two recently published glaciological models of the GIS: i) a model tuned to fit both RSL data from a distribution of sites across Greenland (Tarasov and Peltier, 2002) and, ii) a model tuned to fit observations of ice extent only (Huybrechts, 2002). These two models are referred to, respectively, as GrB and Hu1 in the following discussion. GrB was tuned assuming the Earth viscosity model VM2 with a 90 km thick lithosphere (Tarasov and Peltier, 2002) and represents the Greenland component of the global ice model ICE-5G (Peltier, 2004). We note that GrB was not calibrated to RSL constraints in the southeast sector of the ice sheet (between Scoresby Sund and the southern tip of Greenland), and that Hu1 is tuned here to observational data from core and seismic evidence from the continental margin (Solheim et al., 1998) that places the LGM ice extent c. 200 km from the present-day coast. To make a direct comparison of the results for GrB and Hu1, we start our analysis by adopting a common Earth and (non-Greenland or background) ice model – ICE5G and the VM2 viscosity model – to generate the RSL predictions shown below.

We show in Figs. 5 and 6 snap shots of modelled ice thickness and ice margin changes since 18 k yrs BP for the study area predicted by GrB and Hu1. GrB predicts a maximum ice thickness change since 18 k yrs BP to present of c. 750 m over our study site, with ice thickness changes elsewhere of c. 500 m. The 18 k yrs BP GrB ice sheet margin roughly follows the present-day coast, with the ice margin retreating inland of this after 10 k yrs BP. GrB predicts no significant late Holocene regrowth of the GIS in the study area. In contrast, Hu1 predicts a thicker ice sheet at 18 k yrs BP and maximum thickness changes since this time of c. 1500 m to the southwest of our field site, and between 1000 and 1250 m to the north. The ice sheet in Hu1 reaches the mid to outer shelf at 18 k yrs BP and remains in this position until 15 k yrs BP. Hu1 predicts that between 15 k and 14 k yrs BP, the ice sheet margin retreated up two bedrock troughs that formerly drained the Sermilik and Kangerdlussuaq glaciers respectively (Fig. 7). This is a consequence of the marine parameterisation of Hu1, in which the maximum grounding line of the ice sheet is controlled by water depth. The retreat

Fig. 6. Modelled ice thickness evolution for the GrB (upper panels) and Hu1 (lower panels) model. The blue circle denotes the Ammassalik field study area.
at this time is associated with the rapid sea-level rise due to Meltwater Pulse 1a. Hu1 subsequently retreats to the present coast at c. 10 k yrs BP with no significant readvance of the ice sheet during the late Holocene.

7.2. Relative sea-level predictions

Both GIA models give a poor fit to our new RSL data (Fig. 8). During the early Holocene, GrB predicts that RSL never exceeded 10 m asl, whereas the field observations suggest a local marine limit at c. 69 m asl. The modelled RSL shows a fall between 11 k and 9 k yrs BP, at which time our observations also suggest net RSL fall (although we note the limited data from this interval). GrB predicts a switch from RSL fall to rise after 7 k yrs BP. This records non-Greenland contributions (see Fig. 10 and related discussion) and not the effect of any late Holocene ice sheet regrowth.

Hu1 predicts RSL c. 5 m above present sea-level between 17 k and 14 k yrs BP, significantly lower than the observed elevation of the local marine limit, and a mid Holocene highstand of c. +11 m at 5 kya BP before falling to present. We record no stratigraphic evidence for a mid Holocene highstand of this magnitude in the lake basins we cored. Any such oscillation must have failed to exceed +3.12 m MSL or QT3, the lowest basin we sampled, would have been re-flooded.

8. Discussion

8.1. Ice margin recession and RSL history in southeast Greenland

There are few data with which to compare our new observations and GIA model predictions in southeast Greenland. The only comparable RSL record that uses isolation basin sequences is from the Nanortalik area in southern Greenland (Sparrenbom et al., 2006a,b). The form and timing of RSL fall at Nanortalik is quite different to that from Ammassalik, demonstrating that each has experienced a distinct glacial/déglacial history. This is no surprise, given the location of the two sites; Nanortalik is located at only 60° north and became ice free c. 14 k cal. yrs BP, during the pronounced increase in air temperatures that defined the end of Greenland stadial 2 (GS-2) (Bennike and Björck, 2002). In contrast, our RSL data suggest that the Ammassalik coast remained ice covered until c. 11 k cal. yrs BP.

Our age estimate for ice margin retreat at Ammassalik agrees with the published offshore radiocarbon chronologies from the immediate
south and north of our study area. As we note above, Jennings et al. (2006) locate the ice sheet margin in the Kangerdlussuaq trough area at the shelf edge at 17 k cal. yr BP, with ice retreating to the middle shelf by c. 15 k cal. yrs BP (Jennings et al., 2002). To the south of Ammassalik, Kuijpers et al. (2003) interpret an oxygen isotope spike recorded in a deep water core (1843 m) located beyond the shelf break and dated to c. 21 k radiocarbon yrs BP (core D597-7P) as evidence for initial deglaciation. They also suggest that the central shelf was ice free by 15 k cal. yrs BP, coincident with a warming observed in the Renland ice core (Johnsen et al., 1992). The fjord setting of our study site means that the exact chronology of ice margin retreat from the shelf was likely influenced by local topographic factors and by the dynamics of the ice stream/valley glacier that once occupied the main Ammassalik Fjord. A delay of c. 1 k to 2 k cal. yrs between the retreat of ice from the mid-shelf to the study area is certainly possible. Comparable studies elsewhere in Greenland demonstrate that the lower limit of perched boulders is a good approximation of the altitude of the local marine limit; lakes immediately below the limit contain marine sediments whilst those immediately above do not, and dates for the onset of organic accumulation in lakes above the marine limit agree with the predicted age of the marine limit based on well-constrained trends in early and mid Holocene RSL (Long and Roberts, 2003; Long et al., 2003, 2006). In addition, we note that the results of an independent dating study that employs surface exposure dating completed to the southeast of Ammassalik (Roberts et al., in press) establishes the age of ice margin retreat from the coast to c. 11 k to 12 k yrs BP, close to that implied by our lakes chronology from above the marine limit here. In summary, we have no reason to doubt the field data and their interpretation relating to the timing and altitude of the marine limit.

8.2. GIA models and RSL observations

We note above a significant mismatch between our RSL observations and the GIA model predictions and now consider possible reasons for this mismatch.

8.2.1. The field data

The lack of isolation basins between the marine limit and c. 23 m asl is unfortunate. It means that our interpretation of the trend in early Holocene RSL is crude and it places considerable weight on our ability to accurately identify the marine limit. Our RSL reconstruction assumes that the minimum ages for the lakes above the local marine limit approximate the age of the marine limit at c. 69 m asl. If our field observations are wrong, or if we have misinterpreted the data, then a closer match between the RSL data and GrB and Hu1 might result. Our approach to identifying and dating the marine limit is similar to that which we have used elsewhere in Greenland, and which we have found to be robust based on morphological, sedimentological and chronological grounds. Our previous RSL work in west Greenland (Disko Bugt) demonstrates that the lower limit of perched boulders is a good approximation of the altitude of the local marine limit; lakes immediately below the limit contain marine sediments whilst those immediately above do not, and dates for the onset of organic accumulation in lakes above the marine limit agree with the predicted age of the marine limit based on well-constrained trends in early and mid Holocene RSL (Long and Roberts, 2003; Long et al., 2003, 2006). In addition, we note that the results of an independent dating study that employs surface exposure dating completed to the southeast of Ammassalik (Roberts et al., in press) establishes the age of ice margin retreat from the coast to c. 11 k to 12 k yrs BP, close to that implied by our lakes chronology from above the marine limit here. In summary, we have no reason to doubt the field data and their interpretation relating to the timing and altitude of the marine limit.

8.2.2. The Earth model

Previous studies in Greenland have demonstrated how RSL predictions are sensitive to variations in lithospheric thickness and Earth viscosity (Tarasov and Peltier, 2002; Fleming and Lambeck, 2004). To assess these factors, we conduct a preliminary sensitivity test to explore the influence of Earth structure on sea-level predictions for the Hu1 model (Fig. 9). We conduct a radial viscosity model, depth parameterised to give an elastic lithosphere; an upper mantle...
bounded by the base of the lithosphere (a variable parameter) and the seismic velocity discontinuity at 670 km depth; and a lower mantle continuing below 670 km to the core–mantle boundary. We define a reference Earth model, which has intermediate values for lithospheric thickness, upper mantle viscosity and lower mantle viscosity. These values are, respectively: 96 km, 5 \times 10^{26} \text{ Pa s} and 10^{22} \text{ Pa s}. Each of these Earth parameters is varied over a wide range (see caption of Fig. 9).

In all cases the resulting sea-level predictions fail to envelope all of the data and there are significant data-model misfits both during the initial period of deglaciation and the Holocene. The adopted ice model fails to capture the magnitude of RSL fall for the range of Earth models considered. A preliminary interpretation of this result is that there is insufficient loss of ice mass to generate the required RSL fall. The magnitude of RSL fall can be amplified by thinning the model lithospheric thickness (Fig. 9a), but even the relatively thin value of 71 km does not approach the magnitude required by the data. We note that a recent Rayleigh wave tomography survey concludes that there is significant variation in seismic lithospheric thickness across Greenland, with the lowest values inferred in the southeast (Darbyshire et al., 2004). However, it is unlikely, given the range of parameters considered in Fig. 9, that spatial variations in Earth model parameters alone will resolve the mismatch between our observations and the model predictions.

8.2.3. The ice model

Other possible causes for the poor fit between the models and our observations include the background (non-Greenland) ice model and/or the details of the Greenland ice model. Fig. 10 shows the separate contributions (not including ‘eustatic’) of Greenland ice (Hu1 and GrB) and the background ice model (ICE-5G minus Greenland) to the RSL history of Ammassalik.

The background prediction contributes >30 m of RSL rise between c. 10 k cal. yrs BP and the present-day. This RSL rise is driven, in large part, by the collapse of the forebulge that developed over Greenland due to the former North American and European ice sheets (Fleming and Lambeck, 2004). On comparing the different predictions in Fig. 10, it is clear that the background ice sheets make a significant contribution to the magnitude of the total RSL at Ammassalik. This contribution is due to long wavelength perturbations of both the sea-floor and the sea-surface (dashed line in Fig. 10) as well as through the eustatic (melt) signal. On comparing our predictions (based on the ICE-5G model) to those of Fleming and Lambeck (2004) at a number of locations around Greenland (not shown), we note that there is a significant difference (10s of metres) in the RSL component due to non-local ice. It is therefore plausible that inaccuracies in the background ice model could contribute significantly to the data-model misfit shown in Fig. 8.

Local isostatic effects dominate the Hu1 predictions for Ammassalik at c. 11 ka cal. BP (Fig. 10). This is not an unexpected result since in the near-field of an ice sheet the RSL prediction will tend to be dominated by the solid surface perturbation due to changes in the local ice load. The Hu1 ice thickness is approximately double compared with that in GrB, equivalent to an average extra ice load of c. 750 m. The effect of this extra load is manifest in the prediction of higher RSL values (by up to 60 m at 16 k yrs BP) for Hu1 compared to GrB. The timing of the switch from RSL rise to fall is at 16–14 k and 9 k yrs BP for Hu1 and GrB respectively. This switch relates to the removal of ice load and onset of local ice retreat. If we assume that the mismatch between data and observations is due to an incorrect local ice load history, then both GrB and Hu1 require a greater LGM ice thickness. The magnitude of the misfit between the models and the earliest data points (c. 11 k cal. yrs BP onwards) suggest that the total ice load change between LGM and present-day is insufficient. Further increasing LGM ice thickness over the study area is plausible as neither model extends to reach the shelf edge over the entire southeast region.

Our observations, and the preliminary modelling experiments presented here, favour a thicker ice sheet model that extended well onto the shelf at the last glacial maximum. This is also consistent with the offshore core data from the region. We note that recent research elsewhere in east and south Greenland also support an enlarged ice sheet that extended well onto the continental shelf at the last glacial maximum. For example, Roberts et al. (in press) present cosmogenic ages and trim line mapping data from the outer part of Seramilik Fjord, Ammassalik, that favour a thick ice sheet, whilst in northeast Greenland, Ó Cofaigh et al. (2004) interpret seismic and sedimentological data as evidence for an extensive ice sheet that extended onto the continental shelf and may have reached the shelf edge break. Moreover, in south Greenland, Fleming and Lambeck (2004) demonstrate, using RSL data and geophysical modelling, that the ice sheet must have extended further onto the shelf than thought previously at the last glacial maximum.

9. Conclusions

This paper presents the results of the first relative sea-level investigations to be completed in southeast Greenland and the first appraisal of two recently published glaciological models. Our conclusions are:

i) The local marine limit in our field site is at c. 69 m. We estimate a minimum age for the formation of this limit, and for local ice retreat, of c. 11 k cal. yrs BP. This is in agreement with existing radiocarbon dates from continental shelf cores and also a recent programme of surface exposure dating completed in the study area (Kuijpers et al., 2003; Jennings et al., 2006; Roberts et al., in press).

ii) Four isolation basins define RSL fall from the marine limit to close to present sea-level between c. 11 ka and 6.5 k cal. yrs BP. Although the rate of RSL fall slowed during this period, we suspect that RSL continued to fall below present sea-level for at least part of the mid and late Holocene.

iii) We compare the new RSL data with predictions generated using two recently published glaciological models of the GIS that differ in the amount and timing of ice loading and unloading over our study area. These two GIS models are coupled...
to the same Earth viscosity model and background (non-Greenland) ice model to aid in the data-model comparison. Neither model provides a close fit to the RSL observations.

iv) A preliminary sensitivity study indicates that changes to the modelled Earth viscosity structure alone will not resolve this data-model mismatch and so the mismatch is most likely due to limitations in the two ice models considered. An improved fit would be possible by increasing the local ice thickness of both models.

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