A geological constraint on relative sea level in Marine Isotope Stage 3 in the Larsemann Hills, Lambert Glacier region, East Antarctica (31 366–33 228 cal yr BP)

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1. Introduction

Observations of relative sea-level (RSL) changes constitute one of the primary data sets for constraining the geometry, volume and melt history of past and present glacial masses. These data can be used to determine the loading and rebound of continental plates in response to changes in the weight of a growing or shrinking ice sheet and can be used by modellers to infer, indirectly, the thickness of the overlying ice through time (Lambeck et al., 1998; Lambeck and Chappell, 2001). RSL is the relative vertical displacement between the sea surface and the land over a period of time. It is influenced by changes in both global ocean volume (eustatic changes in sea level) and local vertical movements and deformation of the solid Earth (isostatic changes) that are caused by changes in sea level and local vertical movements and deformation of the solid Earth (isostatic changes) that are caused by changes in the mass of overlying ice and water (Agostinetti et al., 2004). In regions of Antarctica where good RSL field data are available, these data can be used to constrain both local ice and earth parameters and, to a lesser extent, global meltwater influx (Ivins and James, 2005; Bassett et al., 2007). A number of relative sea-level curves that span the Holocene are now available for Antarctica (Zwartz et al., 1998; Hall et al., 2003; Bentley et al., 2005; Verleyen et al., 2005).

In contrast, there is little published geological evidence of RSL that can be used to constrain Antarctic Ice Sheet thickness prior to the Last Glacial Maximum (LGM) (Huybrechts, 2002). This has meant that most estimates of pre-LGM Antarctic ice volume are derived from global models where the Antarctic contribution is largely unconstrained. Instead it is calculated by subtracting the volumes of other global ice sheets (e.g. Greenland) from the global eustatic total (Andrews, 1992). This lack of geological evidence for glacial isostatic adjustment in Antarctica has partly resulted from the assumption that most of the Antarctic coastline was overridden by the Antarctic Ice Sheet at the LGM and therefore the pre-LGM geological evidence was removed by glacial erosion.

The best pre-LGM RSL evidence is in East Antarctica and consists of raised beaches along the margins of the Syowa Coast in Lützow-Holm Bay (Miura et al., 1998a; Nakada et al., 2000, Fig. 1; Anderson et al., 2002). There AMS radiocarbon dating of over 180 in situ fossil molluscs (Laternula elliptica) yields two groups of dates. One group is approximately Holocene in age (c.1030–10 590 14C yr BP) and a second group is mostly older than 30 000 14C yr BP (ranging from c.22 800 to 46 420 14C yr BP) (Miura et al., 1998a). The latter are interpreted as indicating ice-free conditions and a higher relative sea level in Marine Isotope Stage 3 (MIS 3) reaching a maximum during the last interstadial prior to the LGM (Yoshida, 1983; Miura et al., 1998a). Observations of the MIS 3 shell material in the vicinity of East Ongul Island and northern Langhovde show that it is embedded in marine matrix and is neither reworked nor broken...
...there is some debate that the MIS 3 dates might relate to older material being incorporated into the deposits of the LGM (Colhoun, 1991; Denton et al., 1991). Independent tests of the MIS 3 radiocarbon dates using electron spin resonance (ESR) showed that 2 out of 6 samples tested had minimum ages (within the large ESR error) that were within MIS 3, whilst the remaining 4 yielded ages up to MIS 7 (Takada et al., 2003). The ESR dating did not test any of the c. 20 14C AMS dates between 22 800 and 34 65014C yr BP.

Recently, geomorphological and palaeolimnological data have shown that parts of the Larsemann Hills were ice-free during the LGM (Burgess et al., 1997; Hodgson et al., 2001, 2005, 2006, Fig. 1). This raised the possibility that there might be pre-LGM marine deposits in isolation basins there. These are coastal lakes cut off from the sea as RSL falls, or flooded by the sea as RSL rises. Dating the transitions between marine and lacustrine sediments and measuring the height of the sill that formerly connected the lake to the ocean enables an accurate determination of mean sea level at the time that the basin became isolated (Bassett et al., 2007). For example, a 158 cm sediment core from Kirisjes Pond studied by Verleyen et al. (2004b, 2005; Fig. 2) was divided into six stratigraphic zones based on the analysis of sedimentary diatoms and examined for marine and freshwater transitions (Table 1). This revealed that the basal zones of the core (KPI and KP II) were likely to have been deposited under marine conditions in MIS 3. At the time these papers were published there was insufficient microfossil evidence to assign, unequivocally, a marine origin to zones KPI and KP II. The age of the marine shells deposited on the Syowa Coast. In order to test this hypothesis we have re-examined zones KPI and KP II (158–144 cm) in the Kirisjes Pond core at higher resolution than Verleyen et al. (2005), applying new chronological control and a series of biological and biogeochemical analyses to determine if the sediment was deposited in a marine or freshwater environment.

1.1. Site description

The Larsemann Hills is an ice-free polar oasis in Eastern Antarctica, located approximately midway between the eastern extremity of the Amery Ice Shelf and the southern boundary of the Vestfold Hills (Fig. 1). At 50 km², it is one of the larger ice-free oases found along East Antarctica’s 5000 km of coastline. More than 150 freshwater lakes are found in the hills, ranging from small ephemeral ponds to large water bodies such as Progress Lake (10 ha and 38 m deep). Many of the lakes contain sediments of up to 3 m thickness, comprising finely layered remains of freshwater benthic microbial mats, interspersed in some lower altitude coastal lakes by marine sediments resulting from marine transgressions. These sediments are particularly well suited to palaeolimnological studies as there is no significant bioturbation, a limited season of open water when wind-induced mixing might encourage resuspension, and microfossils and biogeochemical markers that are remarkably well-preserved. Kirisjes Pond (69°22’15.83”S, 76°08’34.60”E, Fig. 2) is a 9 m deep freshwater lake located on Kolloy, with a sill height of 8 m above sea level (a.s.l.), based on Australian Antarctic Data Centre GIS (spot-height elevation accuracy of \(\pm 0.5\) m).

2. Methods

Zones KPI and KP II were dated at 150, 155 and 156 cm sediment depth by Accelerator Mass Spectrometry radiocarbon analysis at

<table>
<thead>
<tr>
<th>Zone</th>
<th>Sediment depth (cm)</th>
<th>Zone age range</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>KPII</td>
<td>144–156</td>
<td>c. 13,580–7700 cal BP</td>
<td>Freshwater</td>
</tr>
<tr>
<td>KPI</td>
<td>156–158</td>
<td>c. 22 800–34 65014C yr BP</td>
<td>Marine</td>
</tr>
<tr>
<td>KPIIV</td>
<td>94–112</td>
<td>c. 7700–7080 cal BP</td>
<td>Marine–Freshwater transition</td>
</tr>
<tr>
<td>KPI</td>
<td>88–94</td>
<td>7080 cal BP–present</td>
<td>Marine</td>
</tr>
<tr>
<td>KPI</td>
<td>0–88</td>
<td>7080 cal BP–present</td>
<td>Freshwater</td>
</tr>
</tbody>
</table>

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Beta Analytic, Florida. These supplemented the existing chronological control for the core which consisted of 8 radiocarbon dates and 2 TL dates (Hodgson et al., 2001; Verleyen et al., 2004a). Calibration of conventional radiocarbon ages to calendar years was undertaken using the calibration data set CalPal-2007Hulu (Weninger and Jöris, 2008) using CalPal-2007online (Danzeaglocke et al., 2008).

To determine sediment provenance (marine or freshwater), diatoms were counted at high resolution using the methods described in Verleyen et al. (2004a, 2005). Pigments (chlorophylls, bacteriochlorophylls and carotenoids) were analysed using high performance liquid chromatography–mass spectrometry (HPLC–MS) following methods described in Ains et al. (2001) in concentrated acetone extracts of samples at 150–151, 154–155, 155–156 and 156–157 cm and compared with the pigment content of freshwater sediments in zone KPIII (sample 135–136 cm) and previously published pigment analyses of both freshwater and marine sediments in the remainder of the core (Squier et al., 2002). Further comparison was made with the pigment content of sediment cores and surface sediment samples from other lakes in the region (Hodgson et al., 2004). As an independent test of the provenance of the organic matter, measurements of carbon and nitrogen concentrations (%) from which C/N is derived, and bulk organic carbon isotopic ratios ($\Delta^{13}$Corg) were obtained on selected samples by combustion on a Carlo Erba 1500 online to a VG Triple Trap and Optima dual-inlet mass spectrometer (Mackie et al., 2005; 2007; Lamb et al., 2006). $\Delta^{13}$Corg values were calculated to the VPDB scale using a within-run laboratory standard calibrated against NBS-19 and NBS-22. Analytical reproducibility was normally better than 0.1‰ (2 sigma).

3. Results

Three new AMS radiocarbon dates for zones KPI and KP II constrain the age of this part of the core (150–156 cm) to between 26 650 ± 220 and 28 750 ± 3014C yr BP (31 366–33 228 cal yr BP; Table 2). The dates are in stratigraphic order both within KPI and KPII and when compared with the previous radiocarbon and TL dates for the core, which are also of late MIS 3 age. It is, therefore, likely that this sediment was deposited in situ and is not reworked.

Diatom species composition in KPI and KPII shows that the section of the core between 158 and 146 cm straddles a marine–freshwater transition (Table 3). Samples from 158 to 156 cm are dominated by a marine flora consisting of Tryblionella marginulata, Thalassiosira oestrupii and Fragilariaopsis curta. T. marginulata is characteristic of early stages of transition between marine and lacustrine conditions (contacts) and is found abundantly in a later Holocene marine to freshwater transition zone in this core (Verleyen et al., 2004b, 2005), and in a core from nearby Pup Lagoon (Verleyen et al., 2004a). The marine diatom flora includes one taxon T. oestrupii that is present in earlier Pliocene deposits in the Larsemann Hills (McMinn and Harwood, 1995) and the Vestfold Hills some 120 km distant (Harwood et al., 2000; Whitehead et al., 2001) and F. curta which is found in the Pliocene Sørødal Formation of the Vestfold Hills but not in the Larsemann Hills (Harwood et al., 2000). However, T. oestrupii and F. curta are both abundant in the contemporary marine waters in Prydz Bay (Scott and Marchant, 2005) and in Holocene sediments along the East Antarctic margin (Armand et al., 2005). Although microfossils are rare, at 154 cm a transition period is indicated by the dominance of stomaticysts that are characteristic of marine–freshwater contacts in the region, and the inwash of often fragmented subaerial diatoms (Pinnularia microstauron, Verleyen et al., 2004b). By 148–146 cm the diatom assemblage is characteristic of brackish to freshwater conditions, indicating the near completion of the transition. Absolute diatom abundance remains low compared with the rest of the core due to high amounts of glacially derived inorganic material deposited across the KPI–KPII boundary.

The pigment contents of sediment extracts between 150 and 157 cm are dominated by non-degraded okonene and bacterio-phaeophytin a, with minor quantities of bacteriophaeophytin a derivatives (Fig. 3, Table 4). This is consistent with a dominance of anaerobic purple sulfur bacteria (Chromatiales) such as the genus Thiocapsa which have been found growing in microbial mats under anoxic conditions in marine and brackish environments (Cauvette et al., 1999).

<table>
<thead>
<tr>
<th>Sediment depth (cm)</th>
<th>Sediment zone</th>
<th>Lab. code</th>
<th>Sample material</th>
<th>$^{14}$C Enrichment (%Modern ±1σ)</th>
<th>$^{13}$C Enrichment (%Modern absolute ±1σ)</th>
<th>Conventional Radiocarbon age (years BP ±1σ)</th>
<th>TL age</th>
<th>Carbon content (% by wt.)</th>
<th>$^{13}$CPDB (‰ ±0.1)</th>
<th>Mean calibrated age (CalPal)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>KPII</td>
<td>AA-35717</td>
<td>C</td>
<td>107.35 ±0.51</td>
<td>106.72 ±0.51</td>
<td>Modern</td>
<td>7.4</td>
<td>–15.8</td>
<td>1703 ±66</td>
<td>1319 ±67</td>
</tr>
<tr>
<td>24</td>
<td>KPII</td>
<td>AA-35742</td>
<td>C</td>
<td>80.39 ±0.42</td>
<td>77.12 ±0.43</td>
<td>1755 ± 40</td>
<td>2.8</td>
<td>–19.9</td>
<td>1703 ±66</td>
<td>1319 ±67</td>
</tr>
<tr>
<td>52</td>
<td>KPII</td>
<td>AA-35743</td>
<td>C DE</td>
<td>46.19 ±0.3</td>
<td>41.7 ±0.43</td>
<td>2085 ±45</td>
<td>1.7</td>
<td>–20.0</td>
<td>2062 ±56</td>
<td>1222 ±56</td>
</tr>
<tr>
<td>86</td>
<td>KPII</td>
<td>AA-35744</td>
<td>C DE</td>
<td>37.76 ±0.32</td>
<td>33.4 ±0.26</td>
<td>6205 ± 50</td>
<td>34</td>
<td>–14.6</td>
<td>7113 ±80</td>
<td>5588 ±80</td>
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<tr>
<td>110</td>
<td>KPII</td>
<td>AA-35745</td>
<td>O DE</td>
<td>34.51 ±0.26</td>
<td>30.8 ±0.27</td>
<td>7825 ± 70</td>
<td>11</td>
<td>–23.0</td>
<td>8650 ±105</td>
<td>7233 ±105</td>
</tr>
<tr>
<td>124</td>
<td>KPII</td>
<td>AA-35746</td>
<td>C DE O S</td>
<td>27.4 ± 0.23</td>
<td>24.1 ±0.23</td>
<td>8545 ± 60</td>
<td>76</td>
<td>–17.0</td>
<td>9525 ±29</td>
<td>8082 ±29</td>
</tr>
<tr>
<td>138</td>
<td>KPII</td>
<td>AA-35747</td>
<td>C DE O S</td>
<td>10 400 ± 65</td>
<td>9980 ±0.1</td>
<td>11</td>
<td>16.5</td>
<td>–16.5</td>
<td>12327 ±184</td>
<td>10716 ±184</td>
</tr>
<tr>
<td>148–150</td>
<td>KPII</td>
<td>TL date</td>
<td>O</td>
<td>26 650 ±220</td>
<td>26 650 ±220</td>
<td>35000 ±3500</td>
<td>23.1</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
</tr>
<tr>
<td>150</td>
<td>KPII</td>
<td>Beta-196950</td>
<td>O</td>
<td>154 ±30000</td>
<td>152 ±30000</td>
<td>23000 ±3000</td>
<td>23.1</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
</tr>
<tr>
<td>152–154</td>
<td>KPII</td>
<td>TL date</td>
<td>O</td>
<td>27 000 ± 230</td>
<td>27 000 ± 230</td>
<td>30000 ± 230</td>
<td>23.1</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
<td>31 366 ±355</td>
</tr>
<tr>
<td>155</td>
<td>KPII</td>
<td>Beta-196951</td>
<td>O</td>
<td>28.750 ±300</td>
<td>28.750 ±300</td>
<td>–43 200</td>
<td>0.7</td>
<td>–25.0</td>
<td>33 228 ±458</td>
<td>33 228 ±458</td>
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<tr>
<td>156–158</td>
<td>KPII</td>
<td>CAMS-50376</td>
<td>O S</td>
<td>28.750 ±300</td>
<td>28.750 ±300</td>
<td>–43 200</td>
<td>0.7</td>
<td>–25.0</td>
<td>33 228 ±458</td>
<td>33 228 ±458</td>
</tr>
</tbody>
</table>

Reference data: Dec 1991 atmosphere* 113 ± 0.5

* Carbon content of dried, pre-treated material.
et al., 2004), and have been reported from anoxic coastal marine lagoons with high sulfide concentrations (Caumette, 1986). This pigment composition differs markedly from that in the freshwater layers, which contain a large number of chlorophyll a and b derived components (chlorins, phaeophorbides sterly chlorin esters; Fig. 3, Table 4) and polar carotenoids (e.g. lutein and zeaxanthin (Squier et al., 2002)) typical of eukaryotic primary producers and cyanobacteria, and trace amounts of purpurins that indicate oxygenated conditions at the time of deposition (Naylor and Keely, 1998; Ains et al., 2000). Okenone and bacteriopheophytin a are absent from the freshwater zones elsewhere in the core (Fig. 3 and Squier et al., 2002). Purple sulfur bacteria occur in a variety of settings including anaerobic sulfide-rich freshwater habitats. Accordingly, the presence of okenone does not, in this case, provide an unequivocal indicator of marine conditions.

Analyses of $\delta^{13}C_{\text{org}}$ and C/N were carried out on samples from 156 to 158 in zones KPI and the base of KPII and compared with marine samples from zone KPIV, and freshwater samples from zones KPIII and KPVI (Table 1, Fig. 4). The samples from KPI and the base of KPII clearly share similar $\delta^{13}C_{\text{org}}$ and C/N values to the marine samples from zone KPIV and show markedly different $\delta^{13}C_{\text{org}}$ values to the freshwater samples from zones KPIII and KPVI which, in common with most freshwater lakes in the Larsemann Hills, consist of the remains of benthic microbial mats. The values for $\delta^{13}C_{\text{org}}$ are consistent with a marine source (marine phytoplankton $\delta^{13}C_{\text{org}}$ typically ranges between $-17^{\circ}_{\text{oo}}$ and $-22^{\circ}_{\text{oo}}$ (Maslin and Swann, 2005)) and, combined the $\delta^{13}C_{\text{org}}$ and C/N data clearly separate freshwater from marine sources (cf. Mackie et al., 2007).

4. Discussion

The results presented here are consistent with sediments between 154 and 158 in zones KPI and KPII being laid down under marine conditions with a marine–freshwater transition complete by c. 148–146 cm. In order for this zone to be used as a relative sea-level constraint it is essential to demonstrate that these sediment layers are unequivocally marine, that the stratigraphy is undisturbed, and that the material is not a reworked fossil deposit.

Fossil evidence that the deposit is marine includes the presence of stomatocysts and marine diatoms, including taxa (T. marginulata) that are characteristic of the onset of marine to freshwater transitions in the region. Chemical evidence comes from $\delta^{13}C_{\text{org}}$ and C/N values that overlap with those in the KPIV marine zone and the presence of anoxygenic purple sulfur bacteria whose occurrence is consistent with enhanced levels of microbial production of sulfide as a consequence of sulfate fertilisation via marine waters entering a restricted marginal basin. Coupled with the low relative abundance of freshwater diatoms (with the exception of a few highly eroded central areas of P. microstauron frustules) and diagnostic freshwater pigments (e.g. those commonly found in the freshwater zones KPIII and KPVI (Squier et al., 2002) that are abundant in all local lakes (Hodgson et al., 2004)) it is unlikely that this is a freshwater deposit.

The radiocarbon dates are in stratigraphic order suggesting that the stratigraphy of zones KPI and KPII is undisturbed. Despite the presence of two diatom taxa T. oestroepii and F. curta found in the Pliocene Sørsdal Formation in the Vestfold Hills and the former also being found in the Pliocene deposit in the Larsemann Hills, there are no uniquely Pliocene diatoms in KPI and KPII: both taxa are common in the contemporary marine environment of

<table>
<thead>
<tr>
<th>Diatoms/Chrysophyte cysts</th>
<th>Habitat</th>
<th>KPI 146 cm</th>
<th>KPI 148 cm</th>
<th>KPI 154 cm</th>
<th>KPI 156 cm</th>
<th>KPI 158 cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tryblonella marginulata</td>
<td>Marine – before transition</td>
<td>0</td>
<td>25</td>
<td>0</td>
<td>54</td>
<td>39</td>
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<tr>
<td>Thalassiosira sp1</td>
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<td>Thalassiosira oestrupi</td>
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<td>0</td>
<td>0</td>
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<td>Fragiliopris curta</td>
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<tr>
<td>Amphora cf. costata</td>
<td>Marine</td>
<td>13</td>
<td>25</td>
<td>0</td>
<td>1</td>
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<tr>
<td>Chrysophyte cysts</td>
<td>Transition zones</td>
<td>0</td>
<td>0</td>
<td>80</td>
<td>18</td>
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<tr>
<td>Pinunula microstauron</td>
<td>Subaerial and freshwater</td>
<td>0</td>
<td>0</td>
<td>19</td>
<td>8</td>
<td>6</td>
</tr>
<tr>
<td>Stauroforma inermis</td>
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<td>25</td>
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<td>0</td>
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<td>1</td>
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<tr>
<td>Stauroeopsis ancesps</td>
<td>Freshwater</td>
<td>25</td>
<td>25</td>
<td>0</td>
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<td>1</td>
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<tr>
<td>Pinnomathidium abudans</td>
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<td>25</td>
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<tr>
<td>Melosira sp1</td>
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<td>0</td>
<td>0</td>
<td>4</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 3: Diatom species composition of sediments between 146 and 158 cm in the Kirisjes Pond core (%). Only species recorded at >2% relative abundance are included.
Prydz Bay (Scott and Marchant, 2005). The good preservation of the pigments okenone and bacteriophaeophytin a indicates that this unit is not an older (Pliocene) deposit that has been reworked on account of the pigment degradation that would arise in such conditions. In particular, bacteriophaeophytin a is particularly sensitive to oxidation, forming structurally distinct bacterioviridin derivatives (Wilson et al., 2004).

Collectively the data suggest that between 33 228 and 31752 cal BP Kirisjes Pond was a basin connected to the sea by a small inlet. The basin was likely ice covered for much of the year (at present it is only ice free for 1–2 months in occasional summers) and anoxic, preventing the abundant growth of photoautotrophs with the exception of the purple photosynthetic bacteria. Marine diatoms were a dominant component of the diatom assemblage, but total diatom concentrations are low compared with the rest of the core. It is therefore likely that these diatoms were not growing abundantly in the largely anoxic basin, but were instead washed in via the inlet during periods of local ice melt which would have allowed pulses of marine water to enter the basin unimpeded by ice. By 154 cm the presence of stomatocysts indicates unstable biological conditions, with cyst formation presumably occurring as a result of seasonal or perennial anoxia and salinity fluctuations associated with seasonal snow-melt inputs and the onset of isolation from the sea. By 148 cm (after 31366 cal BP) the diatoms indicate that brackish to freshwater conditions had become established. Increasing inputs of inorganic material suggest that the setting was ice proximal after the transition from zone KPI to KPII, likely discharged from a local ice mass, formed by the accumulation of snow and ice in the catchment such as occupies valley catchments on Storres, 7 km landward of Kirisjes Pond, today. As a result absolute diatom concentrations decline and diatoms are more or less absent between 154 and 146 cm, whereas the pigments of the purple photosynthetic bacteria were detectable up to 150 cm. This suggests that although brackish to freshwater conditions had become established the purple photosynthetic bacteria were able to survive in the basin up to a minimum of 31 366 cal BP after which freshwater conditions were established, as evidenced by the near complete dominance of the freshwater diatom Stauroforma inermis above 145 cm (Verleyen et al., 2004b). The occupation of the catchment by snow and ice resulted in restricted sedimentation in the basin through the LGM, as seen in nearby Progress Lake (Hodgson et al., 2006). This restricted sedimentation persisted until KPIII, 144 cm, extrapolated to c. 13 500 cal BP (Verleyen et al., 2004b), when there is evidence from diatom stratigraphies of continued freshwater conditions and at least seasonal open water as before the second, Holocene, marine transgression (KPIV, 112–94 cm c. 7570–7080 cal yr BP).

The significance of these results is that they demonstrate a marine transgression during MIS 3 at the Larsemann Hills that can be accurately dated. This can be compared with the evidence for an East Antarctic marine transgression inferred from marine fossils and raised beaches along the Syowa Coast in MIS 3 (Maemoku et al., 1997; Miura et al., 1998a) (Fig. 5). There the presence of late Pleistocene beach deposits has been attributed to a warmer period during MIS 3 between 61 000 and 29 000 yr BP, as identified in Antarctic ice cores (Petit et al., 1999) (Fig. 5) and marine sediment cores from the Southern Indian Ocean (Hayes et al., 1976), and is consistent with ice-free conditions inferred from emerged marine deposits along other parts of the East Antarctic coastline (Berkman et al., 1998). The RSL transgression on the Syowa Coast formed...
raised beaches up to c. 10 m (Kai-no-hama and Kominato Beaches, Miura et al., 1998b) and isolated marine deposits up to 20 m (radiocarbon sample Q940126-1, 36 790$^{14}$C yr BP, Miura et al., 1998a) above present sea level. In contrast, the RSL transgression in the Larsemann was at least c. 8 m above present sea level (the sill height of Kirisjes Pond) between 28 750 and 26 650$^{14}$C yr BP (Petit et al., 1999) above present sea level. In contrast, the RSL transgression in the Larsemann Hills is based on data presented in Verleyen et al. (2005). These are presented as uncalibrated ages. MIS 3 marine transgressions in the Larsemann Hills are based on data presented in Verleyen et al. (2005).

The cause of this RSL transgression must involve a combination of local ice mass change combined with changes in global ice volume. At ~35 000 cal BP eustatic sea level (directly related to global ice volume) is estimated to have been at least 60 m lower compared to the present (interglacial value; Fig. 5). It follows that regional ice was more extensive leading to a lowering of the land surface such that both regions (Larsemann Hills and Syowa Coast) were susceptible to a marine transgression at this time. The transgression itself could be associated with either or both: local ice mass changes; or changes in global ice volume. We discuss the evidence for these below.

First, evidence for local ice mass change is supported by geomorphological data from the western Lambert Glacier–Amery Ice Shelf system near Radok Lake which suggest that earlier in MIS 3 grounded ice thickened by at least 100 m adjacent to the Amery Oasis (Goodwin and Hellstrom, 2007). Offshore, studies of a trough mouth fan on the upper continental slope of Prydz Bay suggest that in this part of East Antarctica the maximum grounding line expansions occurred well before the eustatic sea level minima during the late Pleistocene 100 ka glacial cycles (O’Brien et al., 2007). Geophysical surveys of the glacial troughs, and megafaults of the last glacial show ice sheet grounding to the shelf edge in many areas along the East Antarctic margin and dating of marine sediment cores supports a pre-LGM expansion and regional thickening of the East Antarctic Ice Sheet (O’Brien, 1994; O’Brien and Harris, 1996). For example, in Prydz Bay, grounding line moraines along the periphery of the Prydz Bay Channel mark the maximum ice position. This ice persisted there untill an early post LGM retreat after 20 770 and 20 230$^{14}$C yr BP (presented as uncalibrated foraminifera dates but calibrated would be c. 24 700–24 200 cal yr BP; not taking into account any regional reservoir corrections that might be required); with retreat marked by the deposition of an iceberg-rafted diamicton (Domack et al., 1998). The East Antarctic Ice Sheet was however not continuously grounded, but focused into large ice streams, thereby enabling marine water to reach some parts of the current coastline, including the Larsemann Hills and Syowa Coast. This spatially discontinuous and diachronous advance of the ice sheet to the continental shelf break has been previously highlighted from geomorphological and sediment core evidence (Anderson et al., 2002). During MIS 3 the regional ice mass was therefore thick enough to depress the Larsemann Hills region below sea level when global (eustatic) sea level is estimated to have been at least 60 m lower compared to the present (interglacial) value (Fig. 5).

Second, evidence for changes in global ice volume is supported by the observation that the marine transgression was contemporaneous at the Syowa Coast and the Larsemann Hills and matches far field data which document several 10–15 m eustatic sea-level variations during MIS 3; the last overlapping with both the Syowa Coast and the Larsemann Hills data at c. 33 000 cal BP (Yokoyama et al., 2007) (Fig. 5). If these eustatic sea-level variations are the dominant component then their duration might be estimated from some of the marine fossils dated from trenches excavated across

![Fig. 5. MIS 3 marine transgression in the Larsemann Hills based on the presence of marine sediments in Kirisjes Pond compared with ice-volume-equivalent sea level derived from the continental-margin earth model E1 presented in Lambeck et al. (2002, Figure 11), and the Vostok Ice Core $\delta^{18}$O$_{\text{atm}}$ data derived from the World Data Centre for Paleoclimatology Miura et al., 1998b) and isolated marine deposits up to 20 m (radiocarbon sample Q940126-1, 36 790$^{14}$C yr BP, Miura et al., 1998a) above present sea level. In contrast, the RSL transgression in the Larsemann was at least c. 8 m above present sea level (the sill height of Kirisjes Pond) between 28 750 and 26 650$^{14}$C yr BP (Petit et al., 1999) above present sea level. In contrast, the RSL transgression in the Larsemann Hills is based on data presented in Verleyen et al. (2005). These are presented as uncalibrated ages. MIS 3 marine transgressions in the Larsemann Hills are based on data presented in Verleyen et al. (2005).]
the Syowa raised beaches, for example at Langhovde where the majority of marine fossils span from (c. 32 430–46 420 14C yr BP), (Maemoku et al., 1997); not taking into account reports of older ages for some of these fossils inferred from ESR dating (Takada et al., 2003).

We therefore hypothesise the following ice history for the Larsemann Hills based on a preliminary interpretation of our sea-level data: (1) thickening of the local ice masses in the Lambert Galcier–Amery Ice Shelf region during MIS 3 occurred to the extent that local RSL was close to the contemporaneous value, (2) regional ice thickening and/or eustatic sea-level rise peaked at c. 33 000 cal BP and caused a c. 1862 yr marine transgression at Kirisjes Pond in the Larsemann Hills, and a longer transgression, or series of transgressions, in the Syowa region c. 22 800–46 420 14C yr BP (Fig. 5), which equates to c. 25 300 to 48 000 cal BP (although calibrations have a larger error at the older end of this range) (3) parts of the Syowa and the Larsemann Hills coastlines were ice free and recorded the RSL maximum or maxima, (4) the increase in global ice volume at the LGM and/or an early post LGM thinning of ice in the Larsemann Hills region caused a relative sea-level fall in both areas towards the LGM (abandonment of raised beach evidence in Syowa and transition from marine to freshwater sediments in Kirisjes Pond), (5) a high postglacial tuning of global ice sheets manifested in a Holocene marine transgression in the Larsemann Hills and Vestfold Hills (Zwartz et al., 1998; Verleyen et al., 2005) and (6) that in the mid to late Holocene isostatic recovery outpaced the slowing rate of global eustatic sea-level rise and as a result there were no more marine transgressions.

Because they reflect former ice-volume changes, these RSL constraints should help inform a number of important and ongoing debates. These include determining the size of the Antarctic Ice Sheets at and before the LGM, determining the timing and style of deglaciation around the ice sheet margin (Anderson et al., 2002) and sea-level variations during MIS 3 (Yokoyama et al., 2007). Some experiments with glacioisostatic models already have shown a requirement for more extensive ice along the East Antarctic during MIS 3 coast to predict the isostatic component of the sea-level response in the Syowa Coast region (Bassett et al., 2007). Furthermore, a greater regional loading during MIS 3 is consistent with modelled data and geological evidence that indicates a rapid and relatively early deglaciation of the Lambert Glacier region (Verleyen et al., 2004a; Bassett et al., 2007).

5. Conclusions

A glacio marine sediment from 154 to 158 cm depth in a sediment core from Kirisjes Pond provides evidence for a MIS 3 marine transgression of at least 8 m in the Larsemann Hills, and a constraint on relative sea-level change in this region of Antarctica.

The timing of deposition of the marine sediment is constrained by three AMS radiocarbon dates between 150 and 156 cm to between 26 650 ± 220 and 28 750 ± 300 14C yr BP (31 366–33 228 cal yr BP), and by TL dates of late MIS 3 age, and the marine origin is supported by marine diatoms, δ13Corg and C/N values overlapping with marine reference material, an absence of diagnostic freshwater pigments and the unusual presence of pigments of anoxicogenous purple sulfur bacteria.

These results can be compared with the evidence for an East Antarctic marine transgression inferred from raised beaches along the Syowa Coast in MIS 3, and suggest that both the Syowa Coast and the Larsemann Hills had ice-free coastlines at that time.

The evidence for a MIS 3 marine transgression also supports an experiment with a recent glacioisostatic model that has a requirement for more extensive ice along the East Antarctic coast to predict accurately the isostatic component of the sea-level response in the Syowa Coast region.

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