Holocene relative sea-level change and deglaciation on Alexander Island, Antarctic Peninsula, from elevated lake deltas

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Field data constraining the rate and spatial pattern of deglaciation and relative sea-level (RSL) change on the Antarctic Peninsula (AP) are relatively sparse, but are needed to improve regional ice sheet and RSL change models, and contribute to better model predictions of future sea-level rise. We investigated the geomorphology, sedimentology and quartz-fraction single aliquot regeneration optically-stimulated luminescence (SAR-OSL) geochronology of elevated deltas around two epishelf lakes, Ablation Lake (AL) and Moutonnée Lake (ML), Alexander Island, Antarctic Peninsula. AL and ML are dammed by George VI Ice Shelf, and maintain a direct hydraulic connection to the sea; hence, their water levels are controlled by changes in RSL. Our aim was to provide new terrestrial constraints on RSL and deglaciation for the southern AP by comparing the formation processes, age and altitude of the AL and ML deltas with: (1) existing RSL curves for the AP; (2) isostatically-coupled AP ice sheet models, and (3) existing AP deglaciation history and SAR-OSL ages from elevated deltas around the nearby inland Hodgson Lake (HL). Although there was insufficient quartz in the ML samples for SAR-OSL dating, the 4.6 ± 0.4 ka SAR-OSL age of the elevated delta at AL represents the last time active deltas were forming higher than present day lake level, and implies: (1) a fall in RSL of up to 14.4 m since the mid-Holocene in this part of Alexander Island, which is consistent with existing field-based RSL chronologies for the AP; (2) relatively smaller ice masses than suggested by some (but not all) isostatically-coupled ice sheet models since the mid-Holocene, and (3) significant mid-Holocene thinning of the AP ice sheets, which is consistent with regional sediment core data and cosmogenic exposure ages, and the 4.4 ± 0.7 ka SAR-OSL age of the lowermost HL delta.

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1. Introduction

Sea-level rise is currently one of the biggest uncertainties in predictions of future global climate change. The Fourth Assessment Report of the Intergovernmental Panel on Climate Change (Bindoff et al., 2007; IPCC, 2007) suggested that decaying ice sheets may contribute more to future sea-level rise than previously thought, but concluded that current ‘limited understanding of the relevant processes’ (Bindoff et al., 2007 p. 409) makes reliable future predictions challenging (Arthern and Hindmarsh, 2006).

The recent retreat and disintegration of many Antarctic Peninsula (AP) ice shelves (Vaughan, 1993; Rott et al., 1996; Vaughan and Doake, 1996; Vaughan et al., 2003; Cook et al., 2005) has highlighted how the accelerated discharge of glaciers from the Antarctic continent can potentially increase rates of global sea-level rise (De Angelis and Skvarca, 2003; Rignot et al., 2004; Scambos et al., 2004; Rignot et al., 2005; Arthern and Hindmarsh, 2006; Vaughan, 2006; Pritchard and Vaughan, 2007; Hock et al., 2009). In addition, the regions around large ice sheets are important areas for studying past sea-level change because they often show the greatest changes in relative sea-level (RSL), especially during glacial to interglacial transitions (Huybrechts, 2002; Bassett et al., 2005).

Palaeo-records and isostatically-coupled ice sheet models of post-Last Glacial Maximum (LGM) deglaciation and RSL change have provided a better understanding of the potential causes and consequences of recent ice mass loss on the AP, and additional insights into how sea level might change with further deglaciation of Antarctica (Payne et al., 1989; James and Ivins, 1998; Bentley, 1999; Ingólfsson and Hjort, 1999; Ivins et al., 2000; Bentley et al., 2005a; Bentley et al., 2005b; Ivins and James, 2005; Bassett et al., 2007; Smith et al., 2007a).

Earlier ice sheet modelling experiments suggest deglaciation of the AP to its present day limits occurred between 14 and 6.5 ka
Fig. 1. Location map of (a) Antarctic Peninsula (AP); (b) Ablation Lake (AL), Moutonne Lake (ML), Hodgson Lake (HL) on Alexander Island and other place names mentioned in text; (c) summary geomorphological map superimposed on aerial photo Ablation Point area covering Ablation Lake and Moutonne Lake; and (d) summary geomorphological map of the Citadel Bastion area superimposed on an aerial photo showing the location of Hodgson Lake.
Field constraints on Holocene rates and spatial patterns of deglaciation and RSL change on the AP remain largely sparse, however, and new data are needed to further improve the predictive ability of isostatically-coupled ice sheet models. Therefore, this paper aims, firstly, to establish the formation process, provenance and geochronology of elevated deltas in the catchments of Ablation Lake (AL), Moutonnée Lake (ML), and Hodgson Lake (HL) on Alexander Island, AP; and secondly, to determine whether these can be used to provide chronological constraints on Holocene deglaciation and RSL change for the southern AP.

Elevated deltas, lake terraces, and palaeo-shorelines are common features of lake basins in Antarctica, providing evidence of past changes in shoreline elevation and semi-permanent (non-seasonal) changes in hydrological equilibrium (Hodgson et al., 2004), as well as important constraints on climate change (Hall et al., 2001, 2002), the timing and scale of local and regional deglaciation, and RSL (Hall and Denton, 1999; Hall et al., 2004; Bentley et al., 2005a).

The formation of elevated deltas at AL and ML was probably related to changes in RSL. AL and ML are epishelf lakes formed on the east coast of Alexander Island, at the Ablation Point Massif where George VI Ice Shelf (GVI-IS) dams two marine embayments (Fig. 1). Epishelf lakes are tidal, stratified water bodies with a lower marine layer that extends under the ice-shelf to the ocean and an upper fresh meltwater layer whose maximum thickness is determined by the draught of the ice shelf (Heywood, 1977; Smith et al., 2007b). Current lake levels in AL and ML are controlled by changes in sea level via their sub-ice-shelf hydraulic connection to George VI Sound (Smith et al., 2006). Therefore, elevated delta deposits in the AL and ML catchments likely represent a former sea-level highstand, which was abandoned when RSL fell. If accurately dated, these deltas can provide, firstly, the most southerly field constraints on RSL change on the AP (cf. Bentley et al., 2005a); and, secondly, key field data to constrain AP deglaciation history, and isostatically-coupled ice sheet models.

In contrast, HL is a recently discovered former sub-glacial lake, land-locked and dammed by the Saturn Glacier, with no hydraulic sub-ice-shelf connection to the sea (Hodgson et al., in press-a,b). The formation of elevated deltas at HL is probably controlled by meltwater availability and/or variations in the thickness of the Saturn Glacier ice dam; hence, the post-LGM deglaciation history and formation of elevated deltas at HL provides a field constraint on the style and rate of deglaciation for south-eastern Alexander Island independent of RSL changes. Dating elevated deltas at this site constrains when the lake ice surface was higher than present because the deltas were probably abandoned when the ice over the lake retreated and the lake level fell, possibly discharging water via its palaeo-spillway (Fig. 2c) (Hodgson et al., in press-a,b).

In the absence of organic matter for radiocarbon dating, we used optically stimulated luminescence (OSL) to establish the geochronology of the elevated deltas in this study. OSL is a property of certain ionic minerals such as quartz and feldspar, which can be used to date the time elapsed since the sediment was last exposed to light (known as the ‘resetting’ or ‘bleaching’ event) (Godfrey-Smith et al., 1988; Aitken, 1998). The total radiation dose received by sediments (palaeo-dose) since the last ‘bleaching’ event can be estimated in the laboratory by forming a dose-dependent growth curve from luminescence signal measurements to produce an equivalent dose ($D_e$) value, measured in SI unit gyres (Gy), and represents the absorbed energy of radiation, where $1\text{ Gy} = 1\text{ J kg}^{-1} \cdot \text{m}^{2} \cdot \text{s}^{-2}$. If the environmental dose rate ($D_I$) (measured in mGy a$^{-1}$) during burial can be evaluated reliably from field and laboratory measurements, a luminescence age can be calculated as $D_e/D_I$ (Aitken, 1998). Sediments whose OSL clock has been completely reset by exposure to infra-red, visible and/or ultra violet electromagnetic radiation are said to be ‘fully-bleached’. Sediments that are incompletely reset are said to be ‘partially bleached’ and provide unreliable depositional ages (Spencer and Owen, 2004).

1.2. Post-LGM palaeoenvironmental change on Alexander Island

Deglaciation from LGM limits, which began as far back as c. 18 ka, accelerated during the early Holocene when ice rapidly retreated from AP continental shelf areas in and around Marguerite Bay (Ingólfsson et al., 1998; Ingólfsson and Hjort, 2002; Ingólfsson et al., 2003; Bentley et al., 2006, 2009 and references therein) (Fig. 1). Bentley et al. (2009) proposed that the general pattern of Holocene climate change on the AP, and, in particular, the two most prominent warm periods of the early and mid-Holocene, were probably driven by a combination of atmospheric forcing (mainly by latitudinal shifts in Southern Westerly wind circulation position and patterns) and oceanographic forcing (in the form of variable Circumpolar Deep Water position and flow rates; e.g., Smith et al., 2007b).

The mid-Holocene period, in particular, was characterised by a period of cross-Peninsula warming c. 5–2 ka (Hodgson et al., 2004). However, most reconstructions of Holocene AP palaeoenvironmental change are based on records located north of Marguerite Bay, which has different physiographic, climatic, glaciological, and geological setting (Fig. 1). Less is known about mid-Holocene to present palaeoenvironmental change on Alexander Island, due, mainly, to a lack of dateable organic material in moraine and core sediments, and concerns about radiocarbon dating of bulk sediments, which might be influenced by ‘old’ carbon (Bentley et al., 2005b; Smith et al., 2007a; Roberts et al., 2008).

Alexander Island is a large (c. 400×80 km), mostly glaciated island situated to the west of the Antarctic Peninsula between 68.7° and 72.6° S (Fig. 1). It is predominantly ice covered with some exposed nunataks and a few notable ice-free areas, such as the Ablation Point Massif. Numerous glaciers flow off Alexander Island; west into the Bach and Wilkins Ice Shelves and Bellingshausen Sea, and east into GVI-IS. GVI-IS is fed both by outlet glaciers from Alexander Island and the ice cap on Palmer Land, an extension of the main West Antarctic Ice Sheet. Cosmogenic surface exposure ages from moraines and exposed bedrock on Alexander Island (Bentley et al., 2006; Hodgson et al., in press-b) suggest significant ice free areas had appeared by the start of the Holocene.

Multi-proxy analysis and radiocarbon dating of lake sediments from ML and MLW within lake catchment moraines from the Ablation Point area (Fig. 1c) show that GVI-IS retreated to at least the Ablation Point Massif area only once in the last c. 11.5 ka. This occurred c. 9600–7500 yr BP, during the early Holocene hypsithermal, after a phase of atmospheric warmth and the intrusion of warm Circumpolar Deep Water onto the continental shelf (Clapperton and Sugden, 1982, 1983; Bentley et al., 2005b; Smith et al., 2006, 2007a,b; Roberts et al., 2008).

Roberts et al. (2008) described the sedimentology of sediment cores from AL and found that the numerous coarse-grained (≤8 mm) clasts found in the main lake core were predominantly of Alexander Island origin. These sediment cores remain undated, however, due to the coarse clastic nature of the sediments and a lack of macrofossils and organic compounds suitable for radiocarbon dating. Bulk surface...
sediment ages are > 10 ka, suggesting a substantial ‘old’ radiocarbon effect exists at this site (Roberts et al., 2008).

At HL, multi-proxy analysis of the lake sediments and hydrological system, and cosmogenic surface exposure ages from the surrounding ice-free ridges, suggest HL was once a sub-glacial lake buried beneath the overriding Last Glacial Maximum (LGM) ice sheet, emerging to become a perennially ice covered lake system sometime during the late Holocene (Hodgson et al., in press-a,b).

2. Site descriptions

2.1. Ablation Lake (AL) (70° 48.974’ S, 68° 27.143 W)

Ablation Lake is located south of Ablation Point at the entrance to Ablation Valley (Figs. 1c and 2a). The geology of the area was mapped by Horne (1968) and the geomorphology by Sugden and Clapperton (1981) and Clapperton and Sugden (1983).

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At AL, GVI-IS is only partially grounded across the entrance of the embayment, creating a continuous hydraulic connection to the seawater in George VI Sound and an in-phase tidal regime between George VI Sound and the lake (Sugden and Clapperton, 1981; Smith et al., 2006). Due to the absence of a continuous bedrock barrier, GVI-IS penetrates into the northern part of the lake basin forming an ice-shelf tongue, which deposits sediments into the lake via an ice conveyor, and into ice push deposits formed by the tidal pressure ridges on the lake shore (Sugden and Clapperton, 1981; Hall et al., 2006; Smith et al., 2006). These deposits can be coarse gravel to boulder in size. Some are of igneous origin, transportedenglacially from the west coast of Palmer Land through GVI-IS, or reworked from well-developed ice-shelf moraines along the coast of Alexander Island (Clapperton and Sugden, 1983; Smith et al., 2007b; Roberts et al., 2008). Two catchment glaciers currently terminate close to the lake.

The elevated delta at the western shore of AL forms a gently dipping 200 m-long by 60 m-wide terrace, c. 14.4 m above present lake level (a.p.l.l.), bisected by the main lake inflow stream at the western rim of the lake.

2.2. Moutonnée Lake (ML) (7° 52.123′ S, 68° 19.635′ W)

Moutonnée Lake is located south of Ablation Point in Moutonnée Valley (Figs. 1c and 2b). The geology and geomorphology of the ML catchment were first described by Horne (1968), Elliott (1974) and Clapperton and Sugden (1983).

The ML catchment bedrock forms part of the Fossil Bluff Group, and, although a small active glacier exists near the head of the valley, larger glaciers have remained relatively stable c. 2 km up-valley for at least 40 years (Fig. 1c) (Horne, 1968; Elliott, 1974; Clapperton and Sugden, 1983; Butterworth, 1991; Browne, 1996). The catchment is composed largely of bedrock to the south, but reworked glacial sediments have been found to the west, in Moutonnée Valley, and north of the main lake (Clapperton and Sugden, 1983).

At ML, GVI-IS is grounded on a partially submerged bedrock sill damming the embayment. However, the lake retains a limited hydraulic connection to the seawater in George VI Sound as revealed by the offset tidal regime that occurs between the lake and George VI Sound (Pearson and Rose, 1983).

Several elevated deltas exist in the ML catchment. The largest two form wide, gently sloping terraces c. 6 m a.p.l.l. at the western shore of ML: one (MLA) is 400 m long by 40 m wide (and has previously been used as an airstrip), the other (MLB) is 150 m long by 50 m.

2.3. Hodgson Lake (HL) (72° 00.549′ S, 66° 27.708′ W)

Hodgson Lake (HL) is located on the southeastern coast of Alexander Island, further inland than AL and ML. HL was first mapped in 2001–2003, and the lake sediments cored in 2002–3 (Hodgson et al., in press-a,b). It is enclosed on three sides by Citadel Bastion and Corner Cliffs and dammed along its northern edge by the Saturn Glacier (Fig. 1d).

The cirque occupied by HL is one of two north-facing cirques abutting the southern margin of the Saturn Glacier adjacent to George VI Sound. The western cirque at nearby Hall Cliff is filled with locally derived ice that abuts the Saturn Glacier. In contrast, the cirque at HL includes a number of ice-free slopes, and most of the ridges at the rim of the cirque are ice-free (Fig. 2c). There are two small remnant catchment glaciers (unofficial names: Corner Cliffs Glacier and Citadel Bastion Glacier, Figs. 1d and 2c) discharging into the southwest corner of the lake.

A spillway for lake outflow exists in the eastern corner, where the lake abuts Corner Cliffs and the Saturn Glacier (Fig. 1c). The thickness of the Saturn Glacier at this point appears to be critical in controlling the level of water in Hodgson Lake: if it were to thin, the lake could drain to a lower level. The lake ice surface is only 2.3–3.9 m above the WGS84 GPS datum. As the maximum measured depth of the lake is 89.9 m this means that the bottom of the lake is c. 86 m below the WGS84 GPS datum (Hodgson et al., in press-a,b).

Along the south shore of HL are two prominent elevated deltas, the surfaces of which are 8.4 and 7.6 m a.p.l.l., and a prominent horizontal break in slope approximately 300 m in length and 13.4 m a.p.l.l. Active deltas are currently forming where small birds-foot deltas extend over the ice at the margin of the lake, and refreezing of meltwater and sediment associated with several ephemeral meltwater streams is creating sedimentary lobes in and around the lake (Hodgson et al., in press-b). Since the present delta is forming as lobes of sediment on top of the present lake ice, the age of the elevated deltas would be expected to increase with height.

3. Methods

3.1. Modern control samples

To test where (and if) sediment was most likely to be exposed to light in the ML hydrological system, we took ‘light-tight’ luminescence dating samples from four modern depositional environments. Samples were taken using copper tubes, sealed at one end, from a shallow depositional inflow stream, a shallow meltwater pond depositing fine sediment, and from the centre of a glaciofluvial gravel deposit. In addition, a sediment trap was left for one year, suspended at a water depth of 47 m in the centre of the ML main basin (Smith et al., 2006).

3.2. Elevated delta samples

The lake catchments, the altitudes of the elevated deltas, and the delta sampling sites were surveyed using an EDM and a Magellan GPS roving unit and, at HL, a high-precision Trimble GPS. Delta profiles were logged in the field and all samples taken from holes excavated for luminescence dating.

Samples for luminescence dating were extracted from elevated deltas in the AB, ML and HL catchments by hammering four copper tubes (22 mm diameter × 20 cm length), sealed at one end, into excavated delta sections that had been cut back at least 0.5 m (Fig. 2d–f). Tubes were removed, sealed and transported frozen in the dark. Bulk sediment was excavated from the same position as the tubes for spectrometric analysis. Tubes were opened under darkroom conditions and 1–2 cm of sediment from the end of the sample, which could have been exposed to light during sampling, was removed from both ends.

The single-aliquot regenerative-dose (SAR) OSL method (Murray and Wintle, 2000, 2003) was used to reconstruct equivalent dose (Dₑ) values from the 90–150 µm or 150–250 µm quartz fraction for modern and elevated delta samples. Infra-red stimulated luminescence (IRSL) was measured at the end of the SAR-OSL cycle to determine whether OSL came from an uncontaminated quartz source (Duller, 2003). Environmental dose rates (Dₑ) were calculated from bulk geochemical composition data obtained by XRF and ICP-MS. Sample preparation procedures, and sampling errors, are described in Tables 1 and 2, and in the online supplementary data. Bulk sedimentological data were obtained following standard procedures (Fig. 3).

4. Results

4.1. Modern control samples

Of the modern catchment samples collected, sufficient quartz for SAR-OSL analysis was extracted from the ML8 Pond sample. Dose rates for ML8 Pond are similar to those from the modern stream and fluvial gravel samples (Table 1), but significantly lower than elevated delta sediments.
Table 1
Summary dose-rate data for elevated delta samples.

<table>
<thead>
<tr>
<th>SUTL no.</th>
<th>Sample environment</th>
<th>Depth (m)</th>
<th>Waterb content (%)</th>
<th>Cosmicc dose rate (mGy a(^{-1}) ± 1(\sigma))</th>
<th>ISGS gamma dose rate (mGy a(^{-1}) ± 1(\sigma))</th>
<th>Dry TSBC(^d) dose rate (mGy a(^{-1}) ± 1(\sigma))</th>
<th>Sample geochemistry(^e)</th>
<th>Dry gamma(^d) dose rate (mGy a(^{-1}) ± 1(\sigma))</th>
<th>Dry betad dose rate (mGy a(^{-1}) ± 1(\sigma))</th>
<th>Mean totalg dose rate (mGy a(^{-1}) ± SE)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1451</td>
<td>ML8 Stream</td>
<td>0</td>
<td>100±5</td>
<td>0.18±0.018</td>
<td>1.603±0.118</td>
<td>1.85</td>
<td>41.9 110 4.10</td>
<td>0.694±0.019</td>
<td>1.488±0.061</td>
<td>1.340±0.044</td>
</tr>
<tr>
<td>1452*</td>
<td>ML8 Pond</td>
<td>0</td>
<td>100±5</td>
<td>0.18±0.018</td>
<td>-</td>
<td>-</td>
<td>1.78 56.9 1.63 5.94</td>
<td>0.828±0.018</td>
<td>1.575±0.059</td>
<td>1.416±0.037</td>
</tr>
<tr>
<td>1453</td>
<td>ML7 Fluvial gravel</td>
<td>0</td>
<td>100±5</td>
<td>0.18±0.018</td>
<td>1.950±0.073</td>
<td>2.29</td>
<td>68.8 1.70 6.31</td>
<td>0.956±0.023</td>
<td>1.931±0.076</td>
<td>1.629±0.044</td>
</tr>
<tr>
<td>1455</td>
<td>ML47(II) Sediment trap(^1)</td>
<td>47</td>
<td>100±5</td>
<td>0.024±0.002</td>
<td>0.873±0.035</td>
<td>0.40</td>
<td>13.3 0.30 0.92</td>
<td>0.158±0.004</td>
<td>0.333±0.013</td>
<td>0.536±0.023</td>
</tr>
<tr>
<td>1456</td>
<td>AB1(A) Elevated delta</td>
<td>14.4(^a)</td>
<td>0.65±0.03</td>
<td>0.142±0.014</td>
<td>0.905±0.006</td>
<td>2.102±0.068</td>
<td>(1) 2.54 42.0 5.77</td>
<td>0.972±0.026</td>
<td>2.057±0.083</td>
<td>3.407±0.039</td>
</tr>
<tr>
<td>1460*</td>
<td>CB6.5 Elevated delta (relict ice/sediment interface)</td>
<td>6.5(^a)</td>
<td>8±1</td>
<td>0.136±0.014</td>
<td>2.237±0.094</td>
<td>(1) 2.29 61.62 1.79 7.80</td>
<td>1.038±0.024</td>
<td>1.982±0.075</td>
<td>2.310±0.065</td>
<td>3.210±0.065</td>
</tr>
<tr>
<td>1464*</td>
<td>CB8.4 Elevated delta</td>
<td>8.4(^a)</td>
<td>16±1</td>
<td>0.192±0.019</td>
<td>2.548±0.056</td>
<td>(1) 2.71 97.16 2.08 9.71</td>
<td>1.247±0.028</td>
<td>2.363±0.090</td>
<td>3.374±0.059</td>
<td>3.740±0.059</td>
</tr>
<tr>
<td>1468*</td>
<td>ML1A(I) Elevated delta</td>
<td>6.0(^a)</td>
<td>5±0.2</td>
<td>0.155±0.015</td>
<td>0.686±0.005</td>
<td>1.698±0.074</td>
<td>(1) 1.93 38.1 1.55 5.28</td>
<td>0.817±0.020</td>
<td>1.636±0.063</td>
<td>2.805±0.036</td>
</tr>
<tr>
<td>1472*</td>
<td>MLA2(I) Elevated delta</td>
<td>6.0(^a)</td>
<td>6±0.3</td>
<td>0.155±0.015</td>
<td>0.642±0.004</td>
<td>1.934±0.068</td>
<td>(1) 2.11 65.0 1.47 6.05</td>
<td>0.882±0.022</td>
<td>1.772±0.070</td>
<td>2.822±0.036</td>
</tr>
<tr>
<td>1476*</td>
<td>ML1B(I) Elevated delta</td>
<td>6.0(^a)</td>
<td>9±0.4</td>
<td>0.155±0.015</td>
<td>2.250±0.053</td>
<td>(1) 2.32 71.5 1.65 6.96</td>
<td>0.988±0.024</td>
<td>1.962±0.077</td>
<td>2.903±0.035</td>
<td>2.903±0.035</td>
</tr>
</tbody>
</table>

\(d\) Depth (cm) below sediment surface; m a.p.l.l. = metres above present lake level.

\(b\) In-situ percentage water content with 10% error; attenuation factors in Aitken (1998) applied to dry dose rate values.

\(c\) K\(_2\)O data was obtained by standard XRF methods and U, Th, Rb by standard ICP-MS methods (described further in Roberts et al. (2008)). (1) = XRF and ICP-MS measurements, potassium as K\(_2\)O, (2) = Laboratory-based gamma spectrometry measurements, potassium as K.

\(d\) Dose rates based on chemical composition and calculated using conversion factors in Adamiec and Aitken (1998). Sediment core gamma dose-rate calculations assume that sediment in a 30 cm sphere surrounding the core is homogenous. Gamma spectrometry dose rate calculations assume Rb at 50 ppm.

\(e\) Sediment core gamma dose-rate calculations assume that sediment in a 30 cm sphere surrounding the core is homogenous. Gamma spectrometry dose rate calculations assume Rb at 50 ppm.

\(f\) The cosmic dose rate was calculated from 0.21 exp[-0.07 (dr) + 0.0005 (dr\(^2\))] mGy a\(^{-1}\) (Prescott and Hutton, 1994) where \(d\) is the current depth of burial and \(r\) is the density of the attenuating medium (\(r\) = 0.998 g cm\(^{-3}\) for water and 2.6 g cm\(^{-3}\) for sediment). For core sediments, \(d\) = depth of water in the lake (89.8 m) + sediment depth and assumes no changes in water/sediment depth throughout burial.

\(g\) Total dose-rate from means of measured beta, gamma and cosmic ray components incorporating external attenuation factors for U, Th and K taken from Medahl (1979) and based on a median grain-size of 100 \(\mu\)m; † = No material of sand fraction or greater (i.e., \(\geq 63 \mu\)m) was present in the ML sediment trap, but we were able to extract sufficient fine-grained sediment for IMS-MS analysis, and calculate a dose rate of 0.536±0.023 mGy a\(^{-1}\). The low dose rate and anomalous bulk sediment geochemistry of this sample are due to encrustation with sea salt; TSBC\(^n\) = 2 except SUTL1464 \(n\) = 4; Palaeodose reconstruction undertaken on asterisked samples. Other samples are included for completeness, but yielded insufficient quartz in the coarse (90–250 \(\mu\)m) fraction for SAR-OSL analysis. Methods are summarised in supplementary data.

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samples. Four discs of purified quartz were made from the ML8 Pond 90–150 µm or 150–250 µm fraction quartz extracts. Two (of four) discs produced insufficient OSL for analysis. The SAR-OSL ages of the two discs that retained a luminescence signal and passed internal validity tests are 7.3 ± 4.2 ka and 9.7 ± 3.4 ka (see Appendix A).

4.2. Elevated delta samples

4.2.1. Ablation Lake (AL)

The elevated delta (AB1) sample at c. 14 m a.p.l.l. consists of predominantly well-sorted sandy gravel sand with sporadic cobbles and small boulders. Overall, the grain size is unimodally distributed, with nearly 55% of the AB1 sample < 2 mm (Fig. 3).

The AB1 OSL sample was taken from the base of the most prominent elevated delta whose surface is c. 14.4 m a.p.l.l. In-situ gamma spectrometry measurements were made in the excavated OSL sample hole. We extracted 50–100 mg of purified 90–250 µm fraction quartz from AB1 for SAR-OSL dating, sufficient for a 16 small aliquot disc SAR-OSL run.

Equivalent dose ($D_e$) values for AB1 were mainly in the 10–20 Gy range and, excluding statistical outliers, unimodally distributed (Fig. 4a). We separated individual ages into groups, defined by which discs passed or failed internal validity tests (recycling ratio—RR in Table 2) using Ward’s method-Euclidean distance cluster analysis, weighted by the individual age errors. Five discs (out of 16) passed internal validity (recycling ratio) tests (Table 2). The weighted mean age of five discs that passed recycling ratio tests is 4.6 ± 0.4 ka, compared to a weighted mean age from all data of 4.9 ± 0.4 ka (Table 2). Three cluster groups (A–C) exist, with a valid age of 4.0 ± 2.5 ka, and valid weighted mean ages of 4.6 ± 0.4 ka and 7.3 ± 2.2 ka, respectively. One outlier (D) has an age of 12 ± 4 ka.

4.2.2. Moutonnée Lake (ML)

Three samples were taken from the two most prominent raised deltas at the western end of ML, c. 6 m a.p.l.l. Two samples (MLA1 and MLA2) were taken 60 m apart. Sample MLA1 (S 70° 51.796, W 068° 21.099) comes from a part of the delta that is composed of sandy gravel deposits, crudely bedded with angular to sub-angular clasts. Sample MLA2 (S 70° 51.797, W 068° 21.166) is from a part of the same delta composed of coarse sand and gravel. The third sample (ML1B; S 70° 51.893, W 068° 21.538) was extracted from MLB, which was composed predominantly of coarse sand and gravel but with a crudely bedded, silty matrix and some cobbles on its upper surface. The MLA1 and MLA2 samples are composed, respectively, of c.73% and c.87% of gravel deposits, crudely bedded with angular to sub-angular clasts. Sample MLA2 (S 70° 51.797, W 068° 21.166) is from a part of the same delta composed of coarse sand and gravel. The third sample (ML1B; S 70° 51.893, W 068° 21.538) was extracted from MLB, which was composed predominantly of coarse sand and gravel but with a crudely bedded, silty matrix and some cobbles on its upper surface. The MLA1 and MLA2 samples are composed, respectively, of c.73% and c.87% of grains > 2 mm (Table 1), and are a well-sorted, unimodally distributed fine to medium sandy gravel (Fig. 3).

Three OSL dating samples (MLA1, MLA2 and ML1B) were taken from the two elevated deltas (6 m a.p.l.l.) at the western end of Moutonnée Lake. In-situ gamma spectrometry measurements made in excavated OSL sample holes for MLA1 and MLA2, but not in ML1B due

<table>
<thead>
<tr>
<th>Cluster group definition</th>
<th>No. of discs</th>
<th>$D_e$ [Gy]</th>
<th>SAR-OSL age [10^3 yr]</th>
<th>Min–Max group range [10^3 yr]</th>
</tr>
</thead>
<tbody>
<tr>
<td>AB1 (SUTL 1452)</td>
<td>All data 15</td>
<td>1.17 ± 0.12</td>
<td>1.08 ± 0.07</td>
<td>1.33 ± 0.70</td>
</tr>
<tr>
<td>All RR passes 5</td>
<td>1.02 ± 0.17</td>
<td>1.24 ± 0.57</td>
<td>16.0 ± 5.3</td>
<td>4.6 ± 0.4</td>
</tr>
<tr>
<td>Group A: all 4</td>
<td>1.26 ± 0.26</td>
<td>1.27 ± 0.97</td>
<td>11.7 ± 8.8</td>
<td>4.3 ± 1.9</td>
</tr>
<tr>
<td>Group B: RR pass 1</td>
<td>1.12 ± 0.27</td>
<td>1.16</td>
<td>13.6 ± 10.8</td>
<td>4.0 ± 2.5</td>
</tr>
<tr>
<td>Group B: all 6</td>
<td>1.10 ± 0.16</td>
<td>1.09 ± 0.41</td>
<td>16.3 ± 4.9</td>
<td>4.6 ± 0.4</td>
</tr>
<tr>
<td>Group B: RR passes 3</td>
<td>1.00 ± 0.19</td>
<td>1.13 ± 0.38</td>
<td>15.8 ± 5.6</td>
<td>4.6 ± 0.4</td>
</tr>
<tr>
<td>Group C: all 4</td>
<td>1.25 ± 0.27</td>
<td>1.35 ± 0.45</td>
<td>24.3 ± 10.5</td>
<td>7.2 ± 1.2</td>
</tr>
<tr>
<td>Group C: RR pass 1</td>
<td>1.10 ± 0.23</td>
<td>1.69</td>
<td>24.8 ± 13.7</td>
<td>7.3 ± 2.2</td>
</tr>
<tr>
<td>Outlier D 1</td>
<td>1.33 ± 0.62</td>
<td>1.06</td>
<td>41.4 ± 17.4</td>
<td>12 ± 4</td>
</tr>
<tr>
<td>CB6.5 SUTL 1460</td>
<td>All data 14</td>
<td>1.08 ± 0.07</td>
<td>1.33 ± 0.70</td>
<td>14.8 ± 1.7</td>
</tr>
<tr>
<td>All RR passes 9</td>
<td>1.04 ± 0.08</td>
<td>1.18 ± 0.44</td>
<td>16.3 ± 2.0</td>
<td>5.1 ± 0.6</td>
</tr>
<tr>
<td>Group A: all 12</td>
<td>1.08 ± 0.08</td>
<td>1.40 ± 0.65</td>
<td>13.2 ± 1.7</td>
<td>4.1 ± 0.5</td>
</tr>
<tr>
<td>Group A: RR passes 8</td>
<td>1.04 ± 0.09</td>
<td>1.25 ± 0.38</td>
<td>14.1 ± 2.1</td>
<td>4.4 ± 0.7</td>
</tr>
<tr>
<td>Outlier B (RR pass) 1</td>
<td>1.04 ± 0.12</td>
<td>0.93</td>
<td>35.8 ± 4.1</td>
<td>11 ± 1</td>
</tr>
<tr>
<td>Outlier C 1</td>
<td>1.07 ± 0.13</td>
<td>0.90</td>
<td>97.0 ± 9.9</td>
<td>30 ± 3</td>
</tr>
<tr>
<td>Group D: all 6</td>
<td>1.03 ± 0.26</td>
<td>0.93</td>
<td>15.7 ± 7.3</td>
<td>58 ± 8</td>
</tr>
<tr>
<td>CB6.4 SUTL 1464</td>
<td>Group A: all 5</td>
<td>–</td>
<td>–</td>
<td>50.7 ± 8.8</td>
</tr>
<tr>
<td>Group B: all 4</td>
<td>–</td>
<td>–</td>
<td>100.2 ± 21.1</td>
<td>30 ± 7</td>
</tr>
<tr>
<td>Outlier C 1</td>
<td>–</td>
<td>–</td>
<td>195.9 ± 26.5</td>
<td>58 ± 8</td>
</tr>
<tr>
<td>Group D: all 5</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>OSLnatural &gt; saturation</td>
</tr>
</tbody>
</table>

The most reliable ages are shown in bold (see Appendix A).

Fig. 3. Grain size profiles (log scale) for the < 2 mm fraction for representative samples from delta samples from Ablation Lake (AL), Moutonnée Lake (ML) and Hodgson Lake (HL). Samples for grain size analysis were split, and dry sieved through a 2 mm mesh. The < 2 mm fraction was treated with 10% HCl (to remove carbonates) and 30% H2O2 (to remove residual organic material) and analysed with a Coulter laser granulometer following standard procedures. The data shown illustrate the relative lack of 90–250 µm sediment in the AL and ML samples compared to the HL delta samples. ML1B (SUTL1476) was not analysed due to insufficient sample volume.

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from the upper (8.4 m) delta sample in the HL catchment. Samples were taken from the 7.6 m a.p.l.l. delta, at c. 6.5 m a.p.l.l. (CB6.5) (buried ice–sediment interface; henceforth referred to as the ‘lower HL delta’) and from the c. 8.4 m a.p.l.l. delta (CB8.4) (near upper delta surface; henceforth referred to as the ‘upper HL delta’). Both samples are composed of light brown medium to coarse sand; 1% of the CB6.5 sample and 27% of the CB8.4 sample are composed of material >2 mm (Fig. 3). The 90–250 µm fraction of the upper and lower delta OSL samples constituted c. 37% and 30%, respectively, of the bulk sediment (Fig. 3). Quartz was rare, but c.100 mg, sufficient for a 16-disk small aliquot SAR-OSL run, was extracted.

$D_e$ values for the lower HL delta were mostly in the 10–20 Gy range and, excluding two statistical outliers, unimodally distributed (Fig. 4b). One main cluster group (A) and two outliers (B, C) were identified (Table 2). Eight samples in Group A that passed validity tests had a weighted mean age of 4.4 ± 0.7 ka, compared to a weighted mean age for all data of 4.6 ± 0.5 ka (Table 2). Two outliers had ages of 11 ± 1 (valid) and 30 ± 3 (not valid) ka.

$D_e$ values and age distributions for the upper HL delta were larger and more dispersed than the lower delta (Fig. 4c; Table 2). Recycling ratio data were not produced for these samples due to an equipment malfunction; hence it was not possible to determine which discs passed internal validation tests. Nevertheless, it was possible to calculate mean age estimates from ten sample runs with equivalent dose errors <25% (Fig. 4b). Weighted Euclidean distance cluster analysis of the $D_e$ data identified three main groups, two (A, B) with age ranges of 11–19 ka and 26–37 ka, respectively, and a ‘beyond OSL saturation’ group (D) of indeterminate age. One outlier (C) has an age of 58 ± 8 ka (Table 2). We discuss the validity and implications of all age data further in the next section.

5. Discussion

5.1. Assessment of OSL data

The quartz fraction was initially chosen for dating analysis because coarser-grained quartz is thought to bleach more quickly than feldspar, and not expected to suffer from anomalous fading (Godfrey-Smith et al., 1988; Aitken, 1998; Wallinga, 2002; Duller, 2003). Unlike quartz, feldspar bleaching rates are not highly dependent on exposure to the ultra-violet component of daylight (Sanderson et al., 2007), which is potentially advantageous in glaciogenic settings and/or high latitude areas with restricted daylight. Preliminary polynuclear TL, OSL, IRSL screening of HL core sediments suggested feldspars deposited in HL were more poorly bleached than quartz (Hodgson et al., in press-a), but further feldspar-based luminescence analyses might be useful, especially for ML samples with insufficient quartz.

During sampling we observed sediment in shallow streams being exposed to strong sunlight while forming currently active deltas, suggesting sufficient exposure to reset the quartz OSL signal exists in this environment, at least during the ‘austral’ summer. Quartz was largely absent in both modern and elevated delta samples, however, and the quartz extracted had a lower sensitivity (OSL per unit dose) compared to other Antarctic environments investigated (e.g., Roberts et al., 2006). Consequently, most aliquots used in this study failed standard SAR-protocol internal quality checks. Fifty-percent or more of the recycling and recuperation data for each sample were not able to extract sufficient quartz from the ML samples for SAR-OSL analysis.

4.2.3. Hodgson Lake (HL)

Samples were taken from a 2 m section of sediment overlying relict lake-ice, exposed where a gully cuts into the lower delta – 100 m from the marginal moraine of the ‘Corner Cliffs Glacier’ (Fig. 2f). The section comprises debris-free transparent relict lake-ice overlain by a unit of crudely laminated fine sands, silts and inter-bedded gravel layers. Isotopic data ($\delta^{18}O$ and $\delta^2H$) confirms that the buried ice is of lacustrine origin, with values more similar to contemporary lake ice than contemporary lake water, and dissimilar to regional values for compressed snow, glaciers and ice shelf ice (Hodgson et al., in press-b).

The topmost 30 cm of the sediment is cryoturbated and is capped by a coarse gravel lag on the upper surface. The uppermost sediment is not glacially-deformed but, as is common in Antarctic environments, the surface sediment has been disturbed, and, therefore, is not suitable for OSL sampling. The interface between the buried relict lake-ice and the overlying sediment is at an altitude of 6.5 m a.p.l.l. The quartz fraction was initially chosen for dating analysis because coarser-grained quartz is thought to bleach more quickly than feldspar, and not expected to suffer from anomalous fading (Godfrey-Smith et al., 1988; Aitken, 1998; Wallinga, 2002; Duller, 2003). Unlike quartz, feldspar bleaching rates are not highly dependent on exposure to the ultra-violet component of daylight (Sanderson et al., 2007), which is potentially advantageous in glaciogenic settings and/or high latitude areas with restricted daylight. Preliminary polynuclear TL, OSL, IRSL screening of HL core sediments suggested feldspars deposited in HL were more poorly bleached than quartz (Hodgson et al., in press-a), but further feldspar-based luminescence analyses might be useful, especially for ML samples with insufficient quartz.

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5.1.1. Modern OSL data

The accuracy of OSL ages in glaciated and polar regions can be compromised where deposition occurred beneath ice and/or during
extended periods of darkness in the austral winter (Doran et al., 1999; Berger and Anderson, 2000; Berger and Doran, 2001; Gore et al., 2001). ‘Old’ ages from modern samples are a persistent problem for OSL dating studies of young (glacio)fluvial sediments (Wallinga, 2002), suggesting that some modern glacial environments do not provide sufficient light exposure to reset the OSL clock to zero.

Meaningful conclusions from our modern samples are limited by the lack of sediment and quartz. The only sample with sufficient quartz, ML8 Pond, was taken from a pool of turbid silt-rich water up to 30 cm deep. If the OSL signal was fully reset by light exposure in a modern environment, we would expect these samples to return zero or near zero OSL ages. However, the turbidity of this pond probably produced differential bleaching of coarser material that had settled out on the bottom of the pond. Therefore, we consider the 7.3 ± 4.2 ka and the 9.7 ± 3.4 ka SAR-OSL ages from two discs (of four), which retained an OSL signal from ML8 Pond to be unreliable as the larger than expected D0 values most likely relate to incomplete bleaching. Furthermore, both discs failed recuperation tests, which suggests changes in luminescence sensitivity and/or thermal transfer during the measurement process (Rhodes and Bailey, 1997; Rhodes, 2000).

The dose rate of our modern pond (ML8) sample of 1.416 ± 0.037 mGy·a⁻¹ does not include any contribution made by the underlying bedrock or take account of potential changes in the rate or nature of sediment deposition over time. Consequently, this dose rate might not be representative of the actual environmental dose rate. Using the mean dose rate of ML delta sediments (2.843 ± 0.036 mGy·a⁻¹) provides a more accurate assessment of the likely environmental dose rate and reduces the ‘zero-age’ offset to c. 4.2 ± 2.4 ka and 5.5 ± 1.9 ka (Appendix A). This age estimate is closer to ‘zero-age’ offsets encountered in other luminescence studies in other parts of Antarctica. For example, Berger and Doran (2001) obtained ‘modern’ ages of between 600 and 3000 years from zero-age testing experiments in Lake Hoare in the McMurdo Dry Valleys of East Antarctica, by applying TL and IRSL dating methods to fine-grained sediments. The other two discs analysed produced insufficient OSL for analysis, consistent with recent bleaching in a modern depositional setting, and our original ‘zero-age’ hypothesis for these samples.

5.1.2. Elevated delta OSL data

Apart from single-grain dating, the SAR-OSL method is the most reliable method currently available for dating glacial sediments (Klasen et al., 2007). Small aliquot samples used are thought to provide the most reliable multiple-grain equivalent dose results from complex depositional environments (Wallinga, 2002; Spencer et al., 2003; Rhodes, 2007). Nevertheless, ambiguity in the interpretation of OSL ages arises from partially bleached sediments since they carry residual signals from transport and depositional cycles earlier in the history of the minerals. Recent OSL dating studies on relict proglacial and glaciofluvial/deltaic sediments have encountered problems with incomplete bleaching of the luminescence signal prior to deposition, leading to anomalously old ages in some samples, and mixtures of differentially-bleached quartz grains (Rendell et al., 1994; Rhodes and Pownall, 1994; Berger and Anderson, 2000; Rhodes, 2000; Bitinas et al., 2001; Spencer and Owen, 2004; Raukas and Stankowski, 2005; Alexanderson and Murray, 2007; Boe et al., 2007; Gemmell et al., 2007; Klasen et al., 2007; Lukas et al., 2007; Thrasher et al., 2009). Fast component OSL quartz grains and grains whose OSL derives from other components could also exist within a deposit that has been unevenly bleached, creating an ‘apparently’ inconsistently bleached sediment (Bailey, 2000; Murray and Wintle, 2003; Singarayer and Bailey, 2003).

Gemmell et al. (2007) suggested that incompletely bleached samples can still provide an approximation of the maximum residual dose, and, where the dose rate can be accurately assessed, a good approximation of a maximum age for the deposit being dated. Moreover, in SAR-OSL analysis, each aliquot produces a ‘date’: hence, the age distribution can be used to assess whether the sample contains a heterogeneous mixture differentially bleached sediment. Analysis of the quartz OSL decay shape can also give some insight into whether partial bleaching might have occurred in more homogeneous cases (Sanderson et al., 2001; Sommerville et al., 2003, 2007).

The age distribution profile of the AL delta sample was relatively well-defined, compared to the HL upper delta age profile (see below), with a weighted mean age of 4.6 ± 0.4 ka for all data that passed internal validation tests (Table 2). However, cluster analysis of the AL delta profile highlights the existence of three possible age-related groups centred on c. 3.7 ka, 5.2 ka, and 7.0 ka (Table 2). This suggests that three stages of well-bleached deposition could have existed (a scenario which is partially backed-up by a disc at c. 7 ka that passed the internal validity tests), and/or incomplete resetting of quartz during the final phase of sedimentary deposition at c. 3–4 ka. The maximum possible age of the AL delta, from our limited dataset, is c. 12 ka; consistent with what is known about the deglaciation history of this part of Alexander Island (see Section 5.2). Where sufficient luminescence sensitivity exists in individual quartz grains, single-grain OSL dating could be used to further characterise the age-profile of the AL delta.

The lower HL delta was the most homogenously bleached sample examined in this study. Aliquots which passed validity tests have a weighted mean age of 4.4 ± 0.7 ka, which is similar to the 4.6 ± 0.5 ka weighted mean age obtained from all aliquots (Table 2). Two outlier ages (c. 11 ka and c. 30 ka) are consistent with age cluster groups from the upper HL delta, suggesting that the lower HL delta sediments could have been derived from the upper HL delta. However, the lower HL delta has a significantly higher C/N ratio of 34.8 compared to the upper HL delta bulk and < 2 mm fractions, which have C/N ratios of 6.7 and 10.4, respectively (Hodgson et al., in press-a). This is probably due to the presence of calcite grains or a source of old carbon in the bulk sediment, suggesting different sediment sources for the upper and lower deltas, each with potentially different bleaching rates and/or bleaching environment.

The wide scatter in the upper HL data probably reflects reworking of the delta deposit resulting in multiple bleeding histories. Therefore, we regard the youngest age of c. 11 ka as a maximum age estimate for the upper delta. Older ages from lower and upper HL deltas probably represent sediment that was incompletely-bleached during the final formation of the upper delta.

5.2. Elevated delta formation, RSL change and deglaciation on Alexander Island

We propose that the AL (and ML) elevated deltas are the end product of glaciofluvial deposition from the catchment and reworking of sediment transported across the lake by lake-ice conveyors (Hall et al., 2006; Smith et al., 2006). After formation, the AL (and ML) elevated deltas were abandoned by changes in local RSL, resulting from LGM deglaciation and isostatic rebound of Alexander Island. The lake level high-stand at AL therefore provides an independent constraint on RSL change.

It seems likely that a large percentage of RSL change and glacio-isostatic rebound on Alexander Island will have been completed by the mid- to late Holocene since, globally, deglaciation-related sea-level rise was effectively complete by the mid-Holocene (Hall and Denton, 1999). Deglaciation and isostatic readjustment models for the AP (based largely on compilations of ‘palaeo’ data) show that the ice sheet was effectively complete by the mid-Holocene (Hall and Denton, 1999). Although models show that most uplift occurs within the first few thousand years following deglaciation (Lambeck, 1993), changes in the AL lake level were probably driven primarily by changes in
isostatic uplift because sea-level remained largely constant from the mid Holocene onwards (Bassett et al., 2007).

The closest available RSL curve from the AP is from Marguerite Bay (Bentley et al., 2005a). The mid-shelf area of Marguerite Bay was deglaciated by c. 12.000 14C yr. BP (c. 11,700–12,100 cal. yr. BP) (Anderson et al., 2002; Ó Cofaigh et al., 2005). The inner part of Marguerite Bay area deglaciated by c. 9000 14C yr. BP. (c. 7800–8400 cal. yr. BP) with relative sea level up to c. 41 m, or more, above present sea level (Bentley et al., 2005a).

Alexander Island and the Marguerite Bay area have different post-LGM deglaciation histories. Isostatic rebound may have been greater on islands in Marguerite Bay, which deglaciated earlier and more fully. Nevertheless, our AL elevated delta RSL age constraint is consistent with the field-based deglaciation and RSL history of Marguerite Bay. The 4.6 ± 0.4 ka constraint on 14.4 m asl of RSL is bracketed by the two most reliable existing constraints on RSL change in the region (Bentley et al., 2005b): the first occurrence of completely freshwater conditions after the transition from marine to terrestrial sediments in cores extracted from uplifted isolation basins on Pourquoi-Pas Island (19.4 m asl at 6420–50 14C yr. BP; c. 5320–5480 cal. yr. BP) (Bentley et al., 2005b) and Horseshoe Island (3.5 m asl at 1320 ± 70 14C yr. BP; c. 600–880 cal. yr. BP) (Wasell and Hakansson, 1992).

Different isostatically-coupled ice sheets models predict different amounts of uplift for the AP (e.g., James and Ivins, 1998; Ivins et al., 2000; Ivins and James, 2005). Predictions using the glacial isostatic adjustment models described in Bassett et al. (2007) and the ICE-5G-VM2 model (90 km-thick lithosphere) (Peltier, 2004) define a range of uplift magnitudes for the late Holocene that are compatible with the observed value of 14.4 m of uplift since c. 4.6 ka. Field constraints on RSL change from elevated deltas on Alexander Island are important, therefore, because they can reduce the number of possible model scenarios, and improve the predictive ability of future isostatically-coupled ice sheet and sea-level change models.

The mid-Holocene formation of elevated deltas at AL is also consistent with the deglaciation history of the Antarctic Peninsula ice sheet since the LGM. Cosmogenic exposure ages and sediment core data suggest: Alexander Island was overridden with ice, over 470 m thick, at the LGM (c. 21 ka) (Bentley et al., 2006); deglaciation of some upper ridges above c. 400 m occurred by c. 13.5 ka BP (Hodgson et al., in press-a); and the present ice sheet configuration existed sometime after c. 7.2 ka (Bentley et al., 2006; Smith et al., 2006, 2007a,b). This timescale of events is in broad agreement with some ice sheet models (e.g., Payne et al., 1989; Huybrechts, 2002), as well as the proposed deglaciation history of HL (Hodgson et al., in press-a,b).

During post-LGM deglaciation, there was a progressive drop in lake level at HL as the Saturn Glacier thinned (Hodgson et al., in press-a,b). To account for the formation of upper and lower deltas, it is likely that both the Saturn Glacier and the HL catchment glaciers remained relatively thick between the exposure of the col until at least 13.5 ka (Hodgson et al., in press-a) during the early Holocene retreat of GVI-LS (Bentley et al., 2005a).

The SAR-OSL age of the upper HL delta suggests deltas were forming on the elevated lake ice surface from c. 11 ka. This is consistent with cosmogenic exposure ages from the HL area, which suggests that the ridge on the western shore of HL emerged sometime after c. 13 ka. The well-defined SAR-OSL age of the lower HL delta places it in the mid-Holocene ‘hypithermal’, a ‘warm period’ widely documented in climate records from the northern Antarctic Peninsula and sub-Antarctic islands (see Hodgson et al. (2004) and Bentley et al. (2009) and for summaries). The age of the lower HL delta implies HL reached its present configuration as a c. 90 m-deep perennially ice covered lake sometime after 4.4 ± 0.7 ka, when the lower delta became isolated from the current hydrological system, and after significant post-LGM and Holocene thinning of AP ice sheets had occurred.

6. Conclusions

(1) We investigated the geomorphology, sedimentology and geochronology of elevated deltas in the two epi-shelf lake catchments of Ablation Lake (AL) and Moutonnée Lake (ML) and compared their formation processes with elevated deltas in the ice-dammed inland Hodgson Lake (HL), providing new terrestrial constraints on relative sea level (RSL) and deglaciation for Alexander Island, Antarctic Peninsula.

(2) Weighted mean SAR-OSL ages that passed internal validity tests from the elevated delta at AL (4.6 ± 0.4 ka) and lower HL delta (4.4 ± 0.7 ka) represent the last time active deltas were forming at higher than present day lake levels in the AL and HL catchments.

(3) SAR-OSL data from the AL elevated delta place a new chronological and vertical constraint on regional RSL change, suggesting uplift to 14.4 m fall in RSL occurred in the last c. 4.6 ± 0.4 ka in this part of Alexander Island. This is broadly consistent with the field-based RSL and Holocene deglaciation history of Alexander Island and the Marguerite Bay region, as well as some glacial isostatic adjustment models (e.g., Peltier, 2004; Bassett et al., 2007), but suggests that some other models might have overestimated ice mass fluctuations and the rate of RSL change across the Antarctic Peninsula since c. 6 ka.

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Appendix A. Supplementary data

CReceived dose data for modern and elevated delta samples AB1, CB6.5 and CB8.4: TD 7/1 is the ratio of test doses 1 and 7; * based on the measured total dose rate value of 1.416 ± 0.037 mGy s−1 (Table 1); # based on the mean dose rate for ML1A, ML2A, and ML1B of 2.843 ± 0.036 (Table 1).

Equivalent dose (D0) method summary: To establish the ages of the deltas and modern samples, we undertook 90–150 or 150–250 μm quartz fraction single aliquot regeneration–optimally stimulated luminescence (SAR-OSL) analysis (Murray and Wintle, 2000, 2003) using modified standard procedures. These included recycling ratio and recuperation measurements to check that sensitivity changes induced by the measurement process are being adequately corrected. Samples were prepared following standard procedures (Sanderson et al., 2001), and analysed using an automated Risø TL/OSL system (TL-DA-15) using blue LEDs with a wavelength of 470 nm for stimulation. After the natural signal was analysed (60 s; 60% power; 125 °C), a sequence of doses and preheats and read-outs were administered to reconstruct the regeneration line. A repeated dose was used to calculate recycling ratios that examine the effectiveness of sensitivity corrections. A normalisation test dose was added, preheated and read out after each dose point read-out. At the end of the run, a final dose was given, followed by an IRSL readout to check for possible feldspar contamination. Sequence
summary: Preheat 220–280 °C for 10 s, groups of 4 discs; readout 60% X-ray fluorescence (XRF) and Inductively Coupled Plasma-Mass Spectrometry (ICP-MS) measurements of elemental concentrations. For XRF, 2–5 g of bulk sediment was hand-crushed using an agate mortar and pestle. Ten major element run were analysed using fused-disc method in a Phillips PW2404 wavelength-dispersive, sequential XRF spectrometer (Norrish and Hutton, 1969). Precision and accuracy are typically 5–10% and were assessed by measuring geostandard BIR1 routinely and quantified as the root mean square deviation (rmsd), expressed in wt.% of the calibration data about the regression line (Govindaraju, 1994; Fitton et al., 1998). ICP-MS analysis of a suite of trace and REE was undertaken on a VG Elemental PQ 2P plasma spectrometer. Triplicate measurements were made on each sample, and bracketed by regular analyses of geostandards BCR, ACV-1, GA, BE-N, DR-N to assess precision and accuracy, which are typically 1–2%. In situ gamma spectrometry was measured from deltas in AL and ML using a ORTEC gamma spectrometer with a 7.62 × 7.62 cm NaI sensor. Count time was 900 s. Laboratory-based gamma spectrometry and TSBC bulk samples were crushed to a fine powder using a steel Tamba mill. TSBC samples were counted for six cycles of 300 s (Sanderson, 1988). Geochemical analyses were performed on a sub-sample of TSBC bulk material. Where applicable, effective total beta dose rate is calculated from water corrected means of TSBC values and chemically derived dry beta dose values. Water contents were measured as received. High resolution, low-level gamma spectroscopy and TSBC bulk material were undertaken on finely powdered bulk sediment with count times of 50,000 s or 100,000 s to determine the concentration of isotopes in the U- and Th-series decay chains and assess the degree of secular disequilibrium.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.geomorph.2009.05.011.

References

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