The Holocene deglaciation of the Byers Peninsula (Livingston Island, Antarctica) based on the dating of lake sedimentary records


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A B S T R A C T

The process of deglaciation in the Antarctic Peninsula region has large implications for the geomorphological and ecological dynamics of the ice-free environments. However, uncertainties still remain regarding the age of deglaciation in many coastal environments, as is the case in the South Shetland Islands. This study focuses on the Byers Peninsula, the largest ice-free area in this archipelago and the one with greatest biodiversity in Antarctica. A complete lacustrine sedimentary sequence was collected from five lakes distributed along a transect from the western coast to the Rotch Dome glacier front: Limnopolar, Chester, Escondido, Cerro Negro and Domo lakes. A multiple dating approach based on 14C, thermoluminescence and tephrachronology was applied to the cores in order to infer the Holocene environmental history and identify the deglaciation chronology in the Byers Peninsula. The onset of the deglaciation started during the Early Holocene in the western fringe of the Byers Peninsula according to the basal dating of Limnopolar Lake (ca. 8.3 cal. ky BP). Glacial retreat gradually exposed the highest parts of the Cerro Negro nunatak in the SE corner of Byers, where Cerro Negro Lake is located; this lake was glacier-free since at least 7.5 ky. During the Mid-Holocene the retreat of the Rotch Dome glacier cleared the central part of the Byers plateau of ice, and Escondido and Chester lakes formed at 6 cal. ky BP and 5.9 ky, respectively. The dating of the basal sediments of Domo Lake suggests that the deglaciation of the current ice-free easternmost part of the Byers Peninsula occurred before 1.8 cal. ky BP.

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1. Introduction

Glacier retreat in the Antarctic Peninsula (AP) region has accelerated over the last several decades in response to increasing air temperatures (Cook et al., 2005; Pritchard and Vaughan, 2007; Cook and Vaughan, 2010; Shepherd et al., 2012). The AP is experiencing one of the fastest changes in glacier mass balances (Navarro et al., 2013; Osmanoglu et al., 2014). The high climate sensitivity of these islands has resulted in multiple glacial advances and retreats in response to climatic fluctuations during the Holocene (Ó Coigheach et al., 2014; The RAISED Consortium, 2014). These Holocene climatic changes have also had consequences for terrestrial ecosystems, including post-glacial rebound and the formation of marine terraces (John and Sugden, 1971; Fretwell et al., 2010; Watcham et al., 2011), permafrost degradation (Oliva and Ruiz-Fernández, 2015), as well as changes in freshwater ecosystems (Toro et al., 2013) and in the distribution of fauna (Sun et al., 2000; del Valle et al., 2002).

Within the current context, the recent glacial retreat recorded during the second half of the 20th century in many areas in the AP region areas are mostly distributed along the coastal fringes and include both nunataks (areas of high relief protruding above the ice sheet) and maritime environments where mean annual temperatures are slightly below 0 °C. Mean annual temperatures in the SSI range between −1 and −2 °C at sea level (Bockheim et al., 2013), and small changes in average temperature and/or precipitation may thus lead to significant changes in glacier mass balances (Navarro et al., 2013; Osmanoglu et al., 2014).

* Corresponding author.
E-mail address: oliva_marc@yahoo.com (M. Oliva).
needs to be framed within the natural pattern of glacial and climatic events that have occurred during the last millennia. Consequently, constraining the age of the deglaciation of present-day ice-free areas in the AP has received renewed attention during recent years. This process of deglaciation started after the Last Glacial Maximum and continued throughout the Holocene (Sugden and Clapperton, 1986; Ingólfsson et al., 1998, 2003; Heroy and Anderson, 2005; Simms et al., 2011; Ó Cofaigh et al., 2014). Cosmogenic dating of glacial landforms (Seong et al., 2009; Balco, 2011; Balco et al., 2013) has complemented earlier studies focused on radiocarbon (14C) dating from marine and terrestrial records (Ingólfsson et al., 1998, 2003). Nevertheless, the timing of deglaciation and the resulting geomorphic implications are still poorly constrained in many of these ice-free environments. Byers Peninsula, in the westernmost part of Livingston Island, is one such example where further paleoenvironmental research on the transition between glacial and lacustrine sediments is required (Ó Cofaigh et al., 2014). Although a recent study of the drainage system in the Byers Peninsula proposed that three different glacial centres existed within the Byers Peninsula during the Holocene (Mink et al., 2014), until now absolute ages of the deglaciation in this peninsula have been based almost solely on 14C ages of the basal sediments from Limnopolar Lake, which reported an estimated age of ca. 8.3 cal ky BP (calibrated thousands of years before present) (Toro et al., 2013), and of several minimum ages from other lake sediment records that ranged between 4 and 5 cal ky BP (Björck et al., 1996).

In this study, we combine absolute and relative dating techniques of lake sediments to clarify the deglaciation process of the Byers Peninsula. The geomorphological distribution of glacial and periglacial landforms provides information that complements the sediment-based geochronology and assists in the interpretation of paleoenvironmental stages. By deriving a chronological framework for five lakes distributed along an east-west transect from the Rotch Dome glacier to the coast of this peninsula, we: (a) improve the limited network of points from which there is chronological information regarding deglaciation, (b) validate the use of radiocarbon (14C) and thermoluminescence dating as complementary absolute dating techniques in Antarctic lakes to determine the onset of lacustrine sedimentation (indicative of the age of lake formation), (c) propose a general model for glacier retreat and landscape evolution in the Byers Peninsula based on geochronological sequences and geomorphological evidence, and (d) compare the evolution of glacier retreat in the Byers Peninsula with regional environmental and climatic proxies.

2. Study area

2.1. Byers peninsula

Livingston Island is the second largest island in the SSI with a surface area of 818 km². The Byers Peninsula is a ca. 60 km² ice-free area - the largest in the SSI - located at its western end. It has been designated an Antarctic Specially Protected Area (ASPA No. 126). Livingston Island constitutes 23% of the area of the SSI archipelago and represents 25% of its total ice volume (Osmanoglu et al., 2014). At present, approximately 697 km² (85% of the surface area) of Livingston Island is covered by glaciers, although the glaciated area has decreased over the last decades: in 1956 glaciers extended over 734 km² (89.7%) (Calvet et al., 1999). Byers Peninsula represents today almost half of the total ice-free area of Livingston Island. On Livingston Island, glacier retreat has been observed over the last decades in both valley glaciers and in the glacier domes that extend across the lowlands. However, the rate of glacier retreat has decelerated between 2002 and 2011 (Navarro et al., 2013), with slightly positive surface mass balances observed from 2007 to 2011 (Osmanoglu et al., 2014). Glacier shrinkage has exposed new land surfaces in the small peninsulas surrounding the Rotch Dome ice cap, including Elephant Point where the new ice-free surface that appeared between 1956 and 2010 represents 17.3% of its 1.16 km² surface (Oliva and Ruiz-Fernández, 2015). However, this pattern has not been observed in the Byers Peninsula, where the frontal moraine lies in contact with the Rotch Dome glacier.

The retreat of the Rotch Dome during the last few millennia has been accompanied by the formation of marine terraces at elevations between 2 and 15 m a.s.l. due to post-glacial rebound (Hall and Perry, 2004). These raised beaches surround the central plateau of the Byers Peninsula (90–140 m), above which several volcanic plugs stand out, such as Start Hill (265 m), Chester Cone (188 m) and Cerro Negro (143 m) (López-Martínez et al., 2012). The only two small ice masses on the Byers Peninsula that currently exist outside the Rotch Dome are located in its NW sector, where the land surface reaches its greatest elevation a.s.l. Numerous islands and lakes have formed in depressions within the hilly landscape of the central plateau of the Byers Peninsula (Toro et al., 2007), five of which are the subject of this study (Fig. 1).

The Byers Peninsula is composed of marine sediments from the Upper Jurassic to Lower Cretaceous and volcanic and volcaniclastic rocks (López-Martínez et al., 1996). The bedrock is heavily weathered and fractured, with extensive evidence of intense frost shattering. As in other ice-free environments of the SSI, periglacial processes and landforms are widespread at all elevations, with the presence of sporadic or discontinuous permafrost up to 20–40 m a.s.l. and continuous permafrost at higher elevations (Vieira et al., 2010; Bockheim et al., 2013). In the central plateau of the Byers Peninsula the permafrost table occurs at 1.35 m depth (de Pablo et al., 2014).

Climate data from 2002 to 2010 show an average annual temperature of −2.8 °C at 70 m a.s.l. and precipitation ranging from 500 to 800 mm (Bahón et al., 2013). Surface runoff is largely restricted to the summer season and is mostly related to snow melt and active layer thawing (Toro et al., 2013). The vegetation cover is moderately abundant in the flat marine terraces, and is largely composed of mosses and grasses (Vera, 2011), with lichens distributed at higher locations. Soils in the Byers Peninsula are shallow and composed of coarse material (Moura et al., 2012).

2.2. Study lakes

The five lakes studied here are distributed across the central plateau of the Byers Peninsula (Fig. 2). They have small areas (<5 ha), shallow depths (1.6–5.3 m at coring sites) and are located at elevations of between 45 and 100 m (Table 1). They are ice-covered except for 2–3 months during the summer.

Chester Lake is the largest lake, located at the foot of Chester Cone, with a maximum depth of 5.3 m. The catchment is characterised by smooth terrain, with the lake covering over 40% of the catchment area. Limnopolar Lake has the largest catchment, which is mostly devoid of vegetation except for small scattered patches of mosses and lichens in wet areas. The maximum depth of Limnopolar Lake is 5.5 m. However, in certain years snow dams may block its outlet during spring thaw before bursting due to melting, leading to short-term rises in lake level of up to 1 m. Escondido Lake is situated in a depression surrounded by three peaks (110–120 m), with a bare and rounded catchment and an irregular shoreline. The lake is deepest (5.3 m) in its eastern part and it shallows to the west. Cerro Negro Lake is located in a small glacial cirque in the upper part of Cerro Negro Hill, between the two summit peaks. The lake is surrounded by steep talus slopes, with fine-grained sediments only in the northern slope. Cerro Negro Lake has the smallest catchment in our study (1.5 ha), and a maximum depth of 4.5 m. Domo Lake is the closest lake to the Rotch Dome, at a distance of only 300 m from the glacial front. It is the shallowest of all the studied lakes with a maximum depth of <2 m. Patterned ground is widespread on the sediments (sands and pebbles) distributed across the Domo Lake catchment, indicating intense cryoturbation processes reworked by the mass-wasting activity.
3. Materials and methods

During field-work in November 2012, we collected the complete sedimentary sequences from four lakes distributed along an approximately east-west transect from the glacier to the coast (Table 1). The lake surfaces were still frozen at the time of coring, enabling the use of lake ice as the drilling platform. The sediment cores were collected from the deepest part of the lakes using a 60 mm UWITEC gravity corer attached to a 6 m-long telescopic aluminium rod to retrieve the unperturbed water–sediment interfaces, while a 90 mm UWITEC piston corer was used in the deepest lakes (Escondido and Chester) to recover their longer sedimentary sequences. The complete sedimentary infill was recovered for all the lakes, although the basal sediments (5–10 cm) of Cerro Negro Lake were lost during core recovery. A total of 22 core sections (9 in Chester Lake, 4 in Escondido Lake, 3 in Cerro Negro Lake and 5 in Domo Lake) were obtained. The complete sedimentary infill of Limnopolar Lake was obtained in 2003 and 2008 coring campaigns (for further details, see Toro et al., 2013). The sediment cores were stored in a cold room at +4 °C prior to subsampling. All cores were split longitudinally and the main visible sedimentological features were described. The cores were visually cross-correlated using the main lithological features (tephras included) with the aim to both characterize the complete
sedimentary infill of the lakes as well as to select the best core to conduct environmental and climate reconstructions.

The age-depth model of each sedimentary sequence was established using two absolute dating techniques (luminescence dating and 14C) combined with a relative dating approach (tephrochronology). A total of 28 aquatic moss macrofossil samples from the sediment cores of Chester (6), Escondido (17) and Cerro Negro (5) lakes were collected for Accelerator Mass Spectrometry (AMS) 14C dating at the Radiochronology Laboratory of the Centre d’Études Nordiques (Laval University, Canada). Antarctic lake sediments have a very low carbon content and aquatic mosses have been found to provide the most reliable dates (Toro et al., 2013). Radiocarbon dating was not possible in Domo Lake sediments due to the lack of moss remnants and very low carbon content in the sediment. Radiocarbon dates were calibrated with the software Calib version 7.0 and the SHcal13 calibration curve (Reimer et al., 2013), selecting the median of the 95.4% distribution (2σ probability interval). The age-depth model for each sedimentary sequence was built using the Bacon R package (Blaauw and Christen, 2011). The construction of the chronological model of Limnopolar Lake record is described in detail in Toro et al. (2013) and in Martínez-Cortizas et al. (2014).

Samples for luminescence dating were taken after opening the sediment cores under red light. Fine sand grains were obtained by sieving and quartz grain extraction was attempted, but no quartz signals were observed with thermoluminescence (TL) and OSL (Optically Stimulated Luminescence) that usually show clear signals when quartz is present. Silt grains between 4 and 11 μm were also prepared for luminescence, but again did not show TL or OSL emission. Both fractions were analysed by X-ray diffraction and weak evidence of quartz was observed, while anorthite, labradorite and weak albite peaks were identified, as well as amorphous materials. Observations under a stereoscopic microscope revealed high glass content. Thus, polycrystal 90–180 μm sand grains were used for dating. Given the presence of feldspars in the samples (anorthite, labradorite and albite) both Infra-Red Stimulated Luminescence (IRSL) and TL were tested, as feldspars usually show luminescence signals under Infra-Red stimulation and characteristic TL glow curves. IR stimulation revealed very weak luminescence signals but TL showed acceptable glow curves for dating without an optical filter. Thus, TL (total bleach method, Aitken, 1985) was finally used for dating. 5 mg aliquots were mounted in stainless steel cups and measured in an automated Risø DA-15 TL/OSL reader system in the luminescence lab of the University of A Coruña (Spain). TL glow-curves were measured with a coupled 9235QA photomultiplier tube (PMT). Laboratory doses were given using a 90Sr/90Y beta source mounted on the reader giving a dose rate of 0.120 ± 0.003 Gy s⁻¹. TL measurements were performed using the total bleach method (Medjahl, 1986) with the residual TL level defined by the TL remaining after 48 h bleaching to sunlight. The TL equivalent doses were calculated using the interval 300–400 °C. The glow curves were determined applying the additive dose method (Aitken, 1985) with linear extrapolation.

The natural dose-rates were estimated from U, Th and K contents using inductively coupled plasma mass spectrometry (ICP-MS) on bulk samples. Conversion factors proposed by Guerin et al. (2011) were applied in this research. For assessing the external alpha contribution an a-

Table 1
<table>
<thead>
<tr>
<th>Lake</th>
<th>Latitude Longitude</th>
<th>Altitude (m asl)</th>
<th>Catchment area (km²)</th>
<th>Lake area (km²)</th>
<th>Coring depth (m)</th>
<th>Coring date</th>
<th>Selected cores for dating (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chester</td>
<td>62°36'41.10&quot; S 61°06'02.28&quot; W</td>
<td>95</td>
<td>0.09</td>
<td>0.037</td>
<td>5.3</td>
<td>18/11/2012</td>
<td>CH12/05-01G (42)</td>
</tr>
<tr>
<td>Escondido</td>
<td>62°37'06.57&quot; S 61°03'36.50&quot; W</td>
<td>92</td>
<td>0.08</td>
<td>0.016</td>
<td>5.2</td>
<td>24/11/2012</td>
<td>CH12/08-01 (122.5)</td>
</tr>
<tr>
<td>Cerro Negro</td>
<td>62°37'47.30&quot; S 61°00'19.99&quot; W</td>
<td>100</td>
<td>0.015</td>
<td>0.0035</td>
<td>2.2</td>
<td>27/11/2012</td>
<td>ES12/03-02 (148)</td>
</tr>
<tr>
<td>Domo</td>
<td>62°37'17.49&quot; S 60°58'32.58&quot; W</td>
<td>45</td>
<td>0.17</td>
<td>0.025</td>
<td>1.6</td>
<td>27/11/2012</td>
<td>ES12/06-01G (59.5)</td>
</tr>
<tr>
<td>Limnopolar</td>
<td>62°37'23.65&quot; S 61°06'23.67&quot; W</td>
<td>65</td>
<td>0.58</td>
<td>0.0221</td>
<td>4.5</td>
<td>30/11/2008</td>
<td>LIM03/1, LIM08/E-D (Toro et al., 2013)</td>
</tr>
</tbody>
</table>
value of 0.02 was added. As water in sediments affects to ionizing radiation effects (and to the dose rate estimate), water saturation of the samples was assumed during burial. The cosmic dose rates were calculated according to Prescott and Hutton (1994) from geographical data (latitude and altitude), sample depth (including depth of the sample in the core and the water column) and density. Estimated cosmic dose rates were below 10% of the total dose rate of most samples.

4. Results

Radiocarbon dates of 28 aquatic moss macrofossil samples from the sediment cores of Chester, Escondido and Cerro Negro lakes are given in Table 2, while the results of TL dating of the deepest possible samples of these three lakes (plus Domo Lake) are shown in Table 3.

4.1. Chester Lake

Three lithological units were defined from the sedimentary records, from oldest to youngest:
- Unit 1 (151–145 cm composite core depth): composed of light brown, fine sand with small rounded pebbles and interpreted as the onset of the sedimentary infill of the lake. A single tephra layer (T3; Liu et al., 2015) was also observed in this unit.
- Unit 2 (145–44 cm composite core depth): composed of millimeter- to centimeter-scale laminated dark brown mosses and light brown clays and silty clays. Two centimeter-thick black tephra layers were present (T1, T2; Liu et al., 2015), interbedded with these laminations.

4.1.1. Chronology

The age model for Chester Lake is shown in Fig. 3. The onset of lake sedimentation is based on the mid-point of the obtained TL age, which roughly agrees with the radiocarbon age located at the same depth. Unit 2 is interpreted as being completely disturbed, with similar folded structures in all the cores retrieved from this lake. Unit 3 has a very low average sedimentation rate of 0.07 mm/yr, except near the two tephra layers where sedimentation rates increased to 2.67 mm/yr.

4.2. Escondido Lake

4.2.1. Lithology

From the bottom to the top, the sedimentary infill of this lake contained four lithological units:
- Unit 1 (156–152 cm composite core depth): a diamicton composed of light bluish clays and rounded centimeter-scale pebbles.
- Unit 2 (152–131 cm composite core depth): composed of millimeter- to centimeter-scale, alternating dark brown mosses layers, light brown clays and silty clays. Several millimeter-thick

Table 2
Results of AMS \(^{14}\text{C}\) dating with samples from the three lakes with living and subfossil mosses.

<table>
<thead>
<tr>
<th>Lake</th>
<th>Core</th>
<th>Lab code</th>
<th>Sample code</th>
<th>Sediment depth (cm)</th>
<th>Combined depth (cm)</th>
<th>(^{13}C)%</th>
<th>Conventional radiocarbon age (yr BP)</th>
<th>Calibrated age (cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chester</td>
<td>Surface sample</td>
<td>ULA-4656</td>
<td>CH04</td>
<td>2.5–2.7</td>
<td>2.6</td>
<td>–26.8</td>
<td>Modern</td>
<td>–</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4242</td>
<td>CH01</td>
<td>10.3–10.8</td>
<td>10.6</td>
<td>–25.4</td>
<td>1125 ± 15</td>
<td>970 ± 20</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4249</td>
<td>CH03</td>
<td>31.7–31.2</td>
<td>31.9</td>
<td>–25.8</td>
<td>3665 ± 20</td>
<td>1000 ± 30</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4244</td>
<td>CH10</td>
<td>3.5</td>
<td>3.5</td>
<td>–27.6</td>
<td>3685 ± 20</td>
<td>3930 ± 50</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4253</td>
<td>CH12</td>
<td>117.0–121.0</td>
<td>146.4</td>
<td>–23.6</td>
<td>4575 ± 20</td>
<td>5955 ± 5</td>
</tr>
<tr>
<td>Escondido</td>
<td>Surface sample</td>
<td>ULA-4657</td>
<td>ES-15</td>
<td>10.8–11.0</td>
<td>10.9</td>
<td>–24.3</td>
<td>1485 ± 20</td>
<td>1330 ± 30</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4650</td>
<td>ES-16</td>
<td>15.9–16.4</td>
<td>16.2</td>
<td>–22.0</td>
<td>1830 ± 20</td>
<td>1715 ± 25</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4651</td>
<td>ES-10</td>
<td>3.0–3.5</td>
<td>11.1</td>
<td>–24.0</td>
<td>1695 ± 15</td>
<td>1455 ± 10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4704</td>
<td>ES-10</td>
<td>16.8</td>
<td>–25.7</td>
<td>1810 ± 20</td>
<td>1655 ± 60</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4701</td>
<td>ES-02</td>
<td>14.6–14.8</td>
<td>22.5</td>
<td>–23.0</td>
<td>2010 ± 15</td>
<td>1920 ± 20</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4702</td>
<td>ES-11</td>
<td>19.0</td>
<td>–22.2</td>
<td>2195 ± 20</td>
<td>2150 ± 50</td>
<td></td>
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<tr>
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<td></td>
<td>ULA-4703</td>
<td>ES-12</td>
<td>23.0–25.3</td>
<td>33</td>
<td>–24.3</td>
<td>2375 ± 20</td>
<td>2350 ± 35</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4655</td>
<td>ES-12</td>
<td>30.0</td>
<td>37.8</td>
<td>–23.5</td>
<td>2595 ± 20</td>
<td>2710 ± 30</td>
</tr>
<tr>
<td></td>
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<td>ULA-4247</td>
<td>ES-13</td>
<td>40.5</td>
<td>43.6</td>
<td>–24.8</td>
<td>2600 ± 20</td>
<td>2720 ± 45</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4653</td>
<td>ES-14</td>
<td>46.4–46.6</td>
<td>54.3</td>
<td>–23.2</td>
<td>3220 ± 20</td>
<td>3405 ± 50</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4652</td>
<td>ES-15</td>
<td>50.5</td>
<td>–25.7</td>
<td>3325 ± 25</td>
<td>3510 ± 70</td>
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<td></td>
<td>ULA-4651</td>
<td>ES-02</td>
<td>64.8</td>
<td>–22.4</td>
<td>3630 ± 15</td>
<td>3890 ± 65</td>
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<tr>
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<td></td>
<td>ULA-4255</td>
<td>ES-04</td>
<td>132.5–124.0</td>
<td>131.6</td>
<td>–24.1</td>
<td>3765 ± 15</td>
<td>4060 ± 130</td>
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<td></td>
<td></td>
<td>ULA-4702</td>
<td>ES-03</td>
<td>130.2–130.5</td>
<td>138.2</td>
<td>–</td>
<td>4265 ± 20</td>
<td>4755 ± 25</td>
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<td>ULA-4243</td>
<td>ES-01</td>
<td>135.0–135.6</td>
<td>143.1</td>
<td>–23.6</td>
<td>4600 ± 20</td>
<td>5165 ± 140</td>
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<tr>
<td>Cerro Negro</td>
<td>Surface sample</td>
<td>ULA-4658</td>
<td>MO-15</td>
<td>10.8–11.0</td>
<td>10.9</td>
<td>–24.3</td>
<td>Modern</td>
<td>–</td>
</tr>
<tr>
<td></td>
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<td>ULA-4246</td>
<td>MO-12</td>
<td>30.0–32.0</td>
<td>43.6</td>
<td>–24.8</td>
<td>2600 ± 20</td>
<td>2720 ± 45</td>
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<td>2720 ± 45</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4248</td>
<td>MO-14</td>
<td>46.4–46.6</td>
<td>54.3</td>
<td>–23.2</td>
<td>3220 ± 20</td>
<td>3405 ± 50</td>
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<td></td>
<td></td>
<td>ULA-4249</td>
<td>MO-15</td>
<td>50.5</td>
<td>–25.7</td>
<td>3325 ± 25</td>
<td>3510 ± 70</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4250</td>
<td>MO-16</td>
<td>64.8</td>
<td>–22.4</td>
<td>3630 ± 15</td>
<td>3890 ± 65</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4251</td>
<td>MO-17</td>
<td>132.5–124.0</td>
<td>131.6</td>
<td>–24.1</td>
<td>3765 ± 15</td>
<td>4060 ± 130</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4252</td>
<td>MO-18</td>
<td>130.2–130.5</td>
<td>138.2</td>
<td>–</td>
<td>4265 ± 20</td>
<td>4755 ± 25</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ULA-4253</td>
<td>MO-19</td>
<td>135.0–135.6</td>
<td>143.1</td>
<td>–23.6</td>
<td>4600 ± 20</td>
<td>5165 ± 140</td>
</tr>
</tbody>
</table>

Table 3
Results of TL dating with samples from the base of the four lakes.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Sediment depth (cm)</th>
<th>Combined depth (cm)</th>
<th>Dose-rate (Gy/ka)</th>
<th>Number of measured aliquots</th>
<th>Equivalent dose (Gy)</th>
<th>Apparent age (ka)</th>
<th>Fading rate (%)</th>
<th>Corrected age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH-1</td>
<td>116.5–122</td>
<td>147.7</td>
<td>1.20 ± 0.00</td>
<td>24</td>
<td>2.98 ± 0.46</td>
<td>2.48 ± 0.43</td>
<td>8.8 ± 3.5</td>
<td>5.90 ± 1.71</td>
</tr>
<tr>
<td>ES-1</td>
<td>144.0–152.0</td>
<td>154</td>
<td>2.18 ± 0.08</td>
<td>4</td>
<td>17.25 ± 0.86</td>
<td>7.93 ± 0.50</td>
<td>8.7 ± 1.6</td>
<td>7.53 ± 2.48</td>
</tr>
<tr>
<td>CN-1</td>
<td>111.0–112.5</td>
<td>112</td>
<td>1.04 ± 0.10</td>
<td>20</td>
<td>3.98 ± 1.15</td>
<td>3.83 ± 1.17</td>
<td>8.5 ± 4.3</td>
<td>2.26 ± 0.72</td>
</tr>
<tr>
<td>DO-1</td>
<td>55.0–58.5</td>
<td>56.8</td>
<td>1.27 ± 0.09</td>
<td>22</td>
<td>1.48 ± 0.16</td>
<td>1.16 ± 0.15</td>
<td>8.5 ± 4.3</td>
<td>2.26 ± 0.72</td>
</tr>
</tbody>
</table>
black tephra layers were interbedded in this sequence.

- Unit 3 (131–66.5 cm composite core depth): composed of laminated, dark grey clays, silty clays and small pebbles with convoluted structures. The centimeter-thick tephra layer T3 was found at the bottom of this unit.

- Unit 4 (66.5 cm–0 cm composite core depth): composed of the same materials as unit 2, and included two centimeter-scale black tephra layers (T1 and T2; Liu et al., 2015).

4.2.2. Chronology

The chronological model of Escondido Lake was based on 17 radiocarbon ages and one luminescence date (Fig. 4). However, the TL date (7.9 ± 0.5 ky) could not be corrected due to the small number of aliquots present in the sample and was rejected due to the significant age difference with the closest radiocarbon date (5165 ± 140 cal. yr BP). The well-constrained part of Unit 2 has a sedimentation rate between 0.1 and 0.12 mm/yr, and since the main lithological features of this radiocarbon-dated section are the same as the lower non-dated one, we estimate the final stage of deglaciation (i.e. top of unit 1) at between 5870 and 6050 cal. yr BP. The convoluted nature of unit 3 precludes any attempt to calculate its sedimentation rate. Unit 4 is quite similar to unit 2, although the mean sedimentation rate is approximately double (0.20 mm/yr). The lowest sedimentation during this period is found in the uppermost 11 cm of the composite sequence, with 0.08 mm/yr.

4.3. Cerro negro Lake

4.3.1. Lithology

The retrieved sedimentary infill of this lake was subdivided in four lithological units:

- Unit 1 (114.5–50.0 cm core depth): composed of laminated to banded white and light and dark greyish clays and silty clays with
undulating structure. A two centimeter-thick black tephra layer (T3) is found at 59.5 cm of core depth.
- Unit 2 (50.0–25.5 cm core depth): made up of faintly banded light clays and silty clays, which dipped at approximately 30°. One millimeter-thick tephra layer was located at 29.5 cm core depth.
- Unit 3 (25.5–4.5 cm core depth): composed of dark clays and two centimeter-thick tephra layers (T1 and T2), with ca. 5 cm of dark brown moss layers and light brown clays at the top of this unit.
- Unit 4 (4.5–0 cm core depth): composed of faintly banded light clays and silty clays with a microbial mat on the surface.

4.3.2. Chronology
Four radiocarbon ages and one luminescence date (Fig. 5) were used to build the chronological model for Cerro Negro Lake. The reliable chronology is limited to the uppermost 42 cm of the sedimentary infill, where the first moss layer has been dated at 3130 ± 45 cal. yr BP. The sedimentation rates are quite variable throughout this core. As in the previous two lakes, the sedimentation rate for the upper 4.5 cm shows the minimum rate of the entire core (0.03 mm/yr), although the rate increases below this layer to 0.16–1.03 mm/yr, being highest near the T1 and the T2 tephras.

4.4. Domo Lake

4.4.1. Lithology
The 59.5 cm of the uppermost sedimentary infill retrieved from this lake is made up of a mixture of dark brown clays, silty-clays, coarse sands and centimeter-scale, rounded volcanic pebbles. The coarse sands that were more abundant in the lowermost 24 cm were interpreted as diamicton, whereas the clays and silts were most abundant in the uppermost 35.5 cm of the core.
4.4.2. Chronology

In Domo Lake there were no living or subfossil mosses, and thus no suitable organic material was present on which to perform radiocarbon dating. Only one luminescence date obtained at the bottom of the core (2.3 ± 0.7 ky) constrains the age of these sediments. The average sedimentation rate for all the sedimentary infill is 0.33 mm/yr, somewhat higher than in the other lakes.

5. Discussion

By combining absolute ages from luminescence and 14C dating with tephrostratigraphic correlations between the lake sequences retrieved from Limnopolar, Chester, Escondido, Cerro Negro and Domo lakes, we have developed a robust chronological framework for the Byers Peninsula. In the following discussion, we use this chronology to evaluate the history of landscape and geomorphic development of the island during the Mid to Late Holocene. We also discuss the additional constraints on the climatic reconstruction provided by geomorphological observations.

5.1. Geochronology of lake sedimentary records in the Byers Peninsula

Lithostratigraphic analysis of the sediment cores from the five studied lakes revealed very different patterns of Holocene sedimentation in the lakes of the Byers Peninsula (Fig. 6). While the sedimentary infills of lakes Limnopolar, Escondido and (less evidently) Cerro Negro exhibited a cm-thick alternating sequence of moss layers and mineral units, the sequence from Domo Lake was composed entirely of homogeneous coarse-grained sediments. Notably, all cores from Chester Lake displayed a highly disturbed post-glacial sedimentary infill with some mosses in the upper part of the sediment cores.

Furthermore, the preservation of several visible tephra layers within all sediment sequences (Liu et al., 2015) enables direct correlations between cores taken from the same lake as well as between those from Limnopolar, Escondido and Cerro Negro lakes. The Chester Lake
sediment cores can only be used for inferring the age of deglaciation, since the cores contained visible evidence of deformation and post-glacial sediments provided inconsistent $^{14}$C dates. It was therefore not possible to establish a robust chronology for the sedimentary infill of this lake. The Domo Lake record showed very little change in either texture or color within the core and lacked organic matter, which, combined with the complete lack of organic remains for radiocarbon dating, restricted the chronological control to the luminescence dating performed on the bottom sediments.

The basal layers in all lakes comprised coarse-grained sediments with presence of decimeter-sized igneous rock fragments. This is interpreted to represent the diamicton deposited during the deglaciation of these lakes, such as was inferred for the nearby Limnopolar Lake (Toro et al., 2013). The diamicton was reached in two lake sequences (Escondido and Domo) and likely near the base in Chester and Cerro Negro lakes. In these two lakes, the basal sediments (5–10 cm) were lost on the lake ice during core recovery but they were observed and their main lithological features were similar to the diamicton recovered from other lakes. Therefore, the basal age for Chester and Cerro Negro lakes is a minimum age estimated for deglaciation, although we suggest that it closely approximates the actual age of ice-free exposure.

Subsequent to the onset of sedimentation in all lakes, there was a gradual change from coarse- to fine-grained particles with inorganic sediments. In Escondido Lake, the first appearance of mosses occurred at around 5.1 cal. ky BP. This pattern was not observed in Cerro Negro Lake, where the base of the core was dated at 7.5 cal. ky BP and the oldest organic sediments started to be deposited at 3.1 cal. ky BP. This discrepancy in the timing of organic deposition in these two lakes may be related to the location of Cerro Negro Lake, which is at a slightly higher elevation than Escondido Lake, and located in a small cirque sheltered by Cerro Negro hill. This topographical setting in the shadow of the hill is very favorable for snow accumulation, as reflected by the longevity of snow patches present today, and is also favorable for greater annual persistence of the lake ice cover. It is possible that between 7.5 and 3.1 cal. ky BP Cerro Negro Lake was mostly ice and snow-covered, thus impeding organic sedimentation. In high latitude lakes extended ice cover reduces the incident radiation absorbed by the lake and limits the time available for moss growth (Lamoureux and Gilbert, 2005) and other primary producers (i.e. algal biofilms).

In Cerro Negro Lake there is an increase in organic matter though the first mosses appeared later in this lake, at 3.1 cal. ky BP. Both ages fit within the mid-Holocene warm period inferred for the northern AP between 4.5 and 2.8 ky BP (Bentley et al., 2009). The presence of mosses in the Byers Peninsula lake sediments has been interpreted to indicate warmer and/or drier conditions (Toro et al., 2013).

The Escondido Lake record shows alternating layers of mosses and clay-silty sediments until 2.3 cal. ky BP, most likely related to changing environmental conditions in the catchment and therefore, in the lake ecosystem. They are probably related to significant variations in the climate regime, with varying moisture and temperature conditions in the Byers Peninsula between 3.8 and 2.3 cal. ky BP. The presence of faintly laminated and inclined clays and silty clays in the Cerro Negro Lake sediments (Unit 3), instead of the well-defined alternation of mosses and clays found in the Escondido Lake, may suggest the dominance of geomorphic processes in the former whereas climate fluctuations played a key role in the latter.

The upper part of the cores spans the last 1.8 cal. ky BP. However, the youngest mosses were dated at 1455 ± 10 and 1330 ± 30 cal. yr BP in Escondido Lake cores and at 1740 ± 10 cal. yr BP in the Cerro Negro Lake
core. The significance of the absence of moss layers in the uppermost sedimentary infill of these two lakes remains unclear. Their presence in the youngest Limnopolar Lake sediments (ca. 0.5 cal. ky BP) suggests that local processes, rather than large-scale climate or environmental fluctuations, may be responsible for this difference. At present, mosses occur in each of these three lakes. Living mosses from the studied lakes have been radiocarbon dated and reported modern ages in all cases, which suggests that no present-day reservoir effect exists in these lakes.

5.2. The process of deglaciation in Byers

The bottom dates of each lake sequence are indicative of the onset of lacustrine sedimentation, which, in turn, suggests the age of deglaciation at each lake. Using paleolimnological and geomorphological evidence, several stages can be inferred for the environmental dynamics of the Byers Peninsula since the last glaciation:

Stage 1 – Deglaciation of the western fringe and highest land (from 8.3 to 5.9 cal. ky BP)

The timing of the onset of the deglaciation of the Byers Peninsula is based on the dating of the oldest moss found in the sediments of Limnopolar Lake, in the west-central part of the peninsula, with a reported age of 7510 ± 80 cal. yr BP (Toro et al., 2013). The chronological model estimated an age for the base of lacustrine sediments of ca. 8.3 cal. ky BP. Therefore, the relatively flat area at the west of Limnopolar Lake was probably the first deglaciated environment apart from the highest nunataks and coastal areas and beaches. The NW section of the Byers Peninsula – the area encompassing the highest lands with elevations exceeding 200 m – likely would have remained glaciated during this time. There is no geomorphological evidence indicating whether the Rotch Dome and the ice that accumulated in the NW section of the Byers Peninsula were linked during this stage. In contrast, the low-altitude terrain in the SW sector would likely have remained ice-free. This suggests that the process of deglaciation started in the W–SW region of the Byers Peninsula, and recession continued towards the higher elevation areas in the eastern sector located above the equilibrium line altitude (Fig. 8).

The oldest age inferred from luminescence dating in our studied lakes was derived in Cerro Negro Lake, at 7.5 ± 2.5 ky. The higher location of Cerro Negro Lake with respect to the plateau suggests that the eastwards retreat of the Rotch Dome was accompanied by a thinning of the ice cap between 8.3 and 7.5 ky. The diminishing ice thickness during the Early Holocene in response to warming climate conditions finally exposed Cerro Negro Lake and lacustrine sedimentation started at 7.5 ky. Nevertheless, the lake could have remained permanently ice-covered for long periods during the following millennia, as previously mentioned. Glacier recession continued and glacial ice disappeared from the slopes descending from Cerro Negro to the plateau, where blockstreams are distributed today. They are currently inactive and heavily lichenized but they may have developed during this phase, suggesting very intense periglacial processes and widespread permafrost conditions in ice-free areas (Ruiz-Fernández and Oliva, in press).

In the SSI, the shrinkage of the ice caps started between ca. 11 and 9.5 cal. ky BP as a consequence of warmer conditions in the AP region, as inferred from ice and marine cores and supported by geomorphological records (Bentley et al., 2009). Glacier retreat was particularly rapid between 10.1 and 8.2 cal. ky BP (Milliken et al., 2009). One of the implications of the loss of glacier volume in the SSI is the exposure of ice-free terrain during the Early Holocene in low-altitude areas of flat topography located distal from ice caps. This is the case of some coastal environments in King George Island, such as Fildes Peninsula that remained ice-free between 11 and 9 cal. ky BP (Watcham et al., 2011) and the edges of Barton Peninsula that became ice-free around 8 cal. ky BP when lakes began to form (Oliva et al., submitted for publication). Moreover, some of the highest lands of the islands with steep slopes were exposed due to the shrinking ice, therefore acting as nunataks. This was observed in Barton Peninsula, where cosmogenic ages suggest deglaciation between 11.9 and 7.6 ky, and in the nearby Weaver Peninsula, where the highest currently ice-free areas were exposed at 8.8 ky (Seong et al., 2009).

It is likely that the same pattern occurred in the Byers Peninsula during this time. The loss of glacier volume of the Rotch Dome would have cleared the western fringe of the peninsula of ice and the glacier thinning exposed a significant part of the highest lands as nunataks. However, no reliable dates are available for the central-west area of the Byers Peninsula. Several lake records collected from this area in the late 1980s and early 1990s reported basal ages ranging from 4 to 5 cal. ky BP (Björck et al., 1996), in all cases showing significant age differences with the records presented in this study (Fig. 7). However, as Björck et al. (1996) conceded, in most cases they did not reach deglacial sediments, and thus their ages do not provide a chronological control for the onset of deglaciation in the Byers Peninsula.

Marine records suggest that warmer conditions continued from 8.2 to 5.9 cal. ky BP, with the Holocene period of minimum sea-ice cover and warm water conditions (Hodgson et al., 2006; Milliken et al., 2009). Consequently, significant glacier thinning and ice margin retreat occurred in many sites in the western AP between 8 and 7 cal. ky BP, as inferred from geomorphological evidence and geochronological data (Bentley et al., 2006). This glacier thinning and retreat consequently increased the rates of isostatic uplift in the SSI (Watcham et al., 2011).

Stage 2 – Deglaciation of the central plateau (from 5.9 ky to 1.8 cal. ky BP)

The Rotch Dome continued to shrink during this phase. There are no geomorphological traces of the position of the glacier front between the deglaciation of Cerro Negro and the deglaciation of Escondido and Chester lakes. The luminescence age of the basal sediments in Chester Lake indicated an age of 5.9 ± 1.7 ky (at 148 cm depth), which is roughly consistent with the oldest dated moss sample in the same core that yielded and age of 5160 ± 120 cal. yr BP (at 146 cm depth). Basal luminescence dating in Escondido Lake provided an apparent age of 7.9 ± 0.5 ky, which we rejected in favor of the projected 14C-AMS age of 5870–6050 cal. yr BP for the base of the core. Two reasons support such rejection: (i) the number of measured aliquots is too small to get a consistent TL age; (ii) there were insufficient grains to perform fading tests to correct the obtained apparent age for this sample. Therefore, assuming the uncertainties of the dates, Chester Lake was deglaciated at around 5.9 ky and Escondido Lake, located 2 km eastwards, at 6 cal. ky BP. Despite their proximity to Limnopolar Lake (i.e., 1.5–3 km), there was a substantial difference of 2.4–2.5 ky in the timing of deglaciation.

The considerable difference in time between Limnopolar, Chester and Cerro Negro lakes may be related to glacier stabilization or with minor glacier advances and retreats following the glacier shrinkage of the ice cap recorded between 8.3 and 7.5 ky. Moreover, in the case of Chester Lake a small glacier centre may have existed during this stage surrounding the Chester Cone (Martínez de Pisón et al., 1996). Such a long period of relative glacier stabilization may have generated moraines. However, the lack of moraine remnants in the area may be associated with the high effectiveness of physical weathering processes in humid Maritime Antarctica that erodes moraines shortly after their formation (Martínez de Pisón et al., 1996). The existence of a partially dismantled moraine at the foot of Cerro Negro might be attributed to this phase, since it shows evidence of a phase of glacier stabilization during the long-term retreat (Ruiz-Fernández and Oliva, in press). The pattern of long-term retreat of the dome-shaped glacier towards the east does not preclude the presence of small glaciers distributed across the peninsula, as was suggested by glacial deposits surrounding the Chester Cone (Martínez de Pisón et al., 1996) and by drainage networks (Mink et al., 2014). However, the location of some of the latest glacier centres suggested in this study is not supported by the existing absolute dates, since one of the proposed last remnants of glacier ice is precisely located at Limnopolar Lake, known to be the first deglaciated lake in the Byers Peninsula at ca. 8.3 ky (Toro et al., 2013). Nevertheless, an isolated
glacier may have persisted over time in the area of Chester Lake, topographically protected at the foot of Chester Cone. An ice cap centred in the uplands of the NW peninsula during this stage is possible, but there is no evidence to assess if it was connected with the Rotch Dome.

In other areas of the SSI, glaciers reached their present limits around 6.1 ky (Watcham et al., 2011; Ó Cofaigh et al., 2014). Following a period with complex climate patterns in the AP region until 4.5 ky, terrestrial and marine records indicate a phase of significant warming between 4.5 and 2.8 cal. ky BP (Mid-Holocene warm period; Bentley et al., 2009). In the Byers Peninsula, the warmest and wettest conditions were recorded between 3.0 and 2.8 cal. ky BP (Björck et al., 1991, 1996). Consequently, the ice cap in the Byers Peninsula must have retreated significantly and by ca. 1.8 cal. ky BP the retreat of the Rotch Dome reached the area surrounding Domo Lake.

### Stage 3 – Deglaciation of the eastern ice free sector of Byers (prior to 1.8 cal. ky BP to present)

Despite the luminescence age of the basal sediments of Domo Lake (2.3 ± 0.7 ky), the fact that tephra layers T1 and T2 do not appear in the sequence suggests that this lake is younger than 1.8 cal. ky BP, when T1 was deposited according to Escondido Lake sediments. Therefore, the area between this lake and the present-day moraine ridge of Rotch Dome (300 m distance) must have become ice-free during the Late Holocene. This date must be considered as approximate and more absolute dates must be obtained in order to completely constrain this last stage of deglaciation.

This area is hilly, with till spread across the landscape and some lakes and ponds filling intra-moraine depressions. The frontal moraine ridge is very steep, showing a significant height (10–25 m) with respect to the deglaciated area. The moraine is composed of unconsolidated gravels and boulders in a sandy matrix, and forms a large primary longitudinal ridge from the north to the south coast of the peninsula. In the central section there are several secondary arches, which may be indicative of different phases of glacial advance or retreat. Based on this geomorphological evidence it is likely that glacier retreat was relatively fast between Domo Lake and its present-day position. Subsequently, the glacier appears to have experienced minor retreats and advances during the Late Holocene that aggregated sediments to the inner slope of the moraine which therefore constitutes a polygenic moraine (Palacios et al., 2015). This pattern of moraine formation hinders the identification of periods with glacier expansion linked to colder and/or wetter conditions during the last two millennia. In other areas of the western AP such as in Rothera Point (Guglielmin et al., 2015) and Palmer Deep (Shevenell and Kennet, 2002) there is evidence of glacial expansion during the last millennium, namely during the Little Ice Age (LIA). Glacial advances during the LIA have been identified in other areas of the SSI (Yoon et al., 2004; Hall, 2007) although in other ice-free environments, such as the Barton Peninsula, glaciers have been stable during the Late Holocene without major significant advances (Oliva et al., submitted for publication). The chronology of these neoglacial advances is not well-constrained (Bentley et al., 2009). Recently, Simms et al. (2012) suggested that the glacial expansion between 1500 and 1700 CE was followed by retreat until 1840 CE. Glacial retreat resulted in an uplift of 2.5 m in the SSI, which was synchronous with the formation of the lowest raised beaches in the Byers Peninsula (Hall and Perry, 2004). In Byers Peninsula some of the ridges of the polygenic moraine may be related to the LIA. Such climate conditions would also be favorable for more abundant long-lasting snow-patches and might have promoted the appearance of small glacier patches in sheltered environments near the highest volcanic plugs, as well as the expansion of the two small glaciers that still exist in the NW corner of the Byers Peninsula.

Finally, the recent accelerated retreat observed in many glaciers of the SSI over the last decades (Navarro et al., 2013) is not clearly visible in the Byers Peninsula. In comparison to neighboring areas where

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**Fig. 7.** Distribution of the available dates from lake sediments in Byers Peninsula, comparing this research with former studies.
Glaciers are now physically disconnected from the moraine system, such as Elephant Point (Oliva and Ruiz-Fernández, 2015), today the Rotch Dome sits in contact with the moraine in the centre of the Byers Peninsula.

6. Conclusions

The analysis of five lacustrine sedimentary sequences (Domo, Cerro Negro, Escondido, Chester and Limnopolar lakes) along an east-west transect from the Rotch Dome glacier to the coast enabled the identification of the past dynamics of the westernmost ice cap in Livingston Island. It provides a continuous record for past environmental and climate changes during most of the Holocene in the western South Shetland Islands. The Holocene retreat of this ice cap has exposed the largest ice-free environment in the SSI, with major consequences for the geomorphological and ecological dynamics in the area. The chronology of the deglaciation in the Byers Peninsula provides a framework for better understanding a wide range of processes connected with glacial retreat in the terrestrial ecosystem, such as colonization of vegetation, distribution of permafrost and geomorphological landforms and processes, formation of marine terraces.

The onset of lake sedimentation is indicative of the age of deglaciation at each site. The retreat of Rotch Dome glacier pre-dated 8.3 cal. ky BP in the westernmost third of the peninsula as a response to warmer climate conditions in the AP region during the Early Holocene. The shrinking of the glacier cleared the highest volcanic plugs of the peninsula of ice, as inferred from Cerro Negro Lake sediments, which indicated that this lake became ice-free by 7.5 ky. The highest lands in the Byers Peninsula therefore acted as nunataks protruding the glacial ice during most of the Mid-Holocene. The loss of glacier volume continued during the Mid-Holocene and the central plateau became gradually ice-free, as indicated by the Escondido and Chester lake records, with these lake catchments becoming ice-free by 6 cal. ky BP and 5.9 ky, respectively. The long-term eastward retreat of the Rotch Dome during the Holocene may have occurred parallel to the existence of other small glacier spots distributed at the foot of the highest elevations of the peninsula. The easternmost ice-free area in Byers became ice-free during the Late Holocene, and probably before 1.8 cal. ky BP, as suggested by Domo Lake sediments.

This research provides new data about the paleoenvironmental history of the Byers Peninsula, and presents new questions for future studies. In particular, the application of cosmogenic dating techniques would complement data about the Holocene deglaciation process inferred from lake sediments by incorporating dates obtained from erosional and depositional glacial evidence. Additional effort should focus on the NW corner of this peninsula, where two small glaciers still exist today, but about which information on past environmental dynamics is lacking.
Acknowledgements

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