Evidence for the long-term sedimentary environment in an Antarctic subglacial lake

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ABSTRACT

Lakes beneath the Antarctic Ice Sheet are of fundamental scientific interest for their ability to contain unique records of ice sheet history and microbial life in their sediments. However, no records of subglacial lake sedimentation have yet been acquired from beneath the interior of the ice sheet, and understanding of sediment pathways, processes and structure in subglacial lake environments remains uncertain. Here we present an analysis of seismic data from Subglacial Lake Ellsworth, showing that the lake bed comprises very fine-grained sediments deposited in a low energy environment, with low water- and sediment-fluxes. Minimum sediment thickness is 6 m, the result of prolonged low sedimentation rates. Based on the few available analogues, we speculate this sediment age range is a minimum of 150 ka, and possibly >1 Ma. Sediment mass movements have occurred, but they are rare and have been buried by subsequent sedimentation. We present a new conceptual model of subglacial lake sedimentation, allowing a framework for evaluating processes in subglacial lake environments, and for determining future lake access locations and interpreting subglacial lake samples.

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1. Introduction

More than 400 lakes lie hidden beneath the Antarctic Ice Sheet (Siegert et al., 2016), part of an extensive active subglacial hydrological system that influences ice dynamics and consequently, the future evolution of the ice sheet (Bentley et al., 2011). Despite being remote and difficult to access, subglacial lakes are of significant scientific interest as repositories for records of ice sheet and microbial history (Christoffersen et al., 2008; Hodgson et al., 2016; Kuhn et al., 2017), and as large sources of water that can lubricate ice motion, impacting ice sheet stability and mass balance (Stearns et al., 2008). At interior ice sheet locations in particular, some subglacial lakes may have remained stable and undisturbed for long periods (>1 Ma) and hence contain longer ice sheet records than the oldest ice core evidence. Beneath the West Antarctic Ice Sheet, subglacial lake sediments may yield evidence for Pleistocene ice sheet collapses. Hence, subglacial lake records may yield ice sheet history and constrain palaeo-ice sheet configurations at locations

and over timescales that cannot otherwise be achieved, and are therefore potentially of unique value in understanding ice sheet-climate interactions.

The presence of sediments is critical to the scientific value of subglacial lakes (Siegert et al., 2012). Successful recovery of ice sheet history and records of microbial communities depends on prolonged periods of undisturbed deposition and on resulting sediments being soft, unconsolidated and poorly lithified (Bentley et al., 2011); however, direct evidence for this is lacking. Although two contemporary subglacial lakes have been accessed, Vostok Subglacial Lake (Lukin and Vasiliev, 2014) beneath the East Antarctic Ice Sheet and Subglacial Lake Whillans (SLW; Tulaczyk et al., 2014) beneath Whillans Ice Stream, no sediments were recovered from the former, and those recovered from SLW reflect ice-stream flow and drainage events, not subglacial lake sedimentation (Hodson et al., 2016). Sedimentary material has been interpreted beneath some subglacial lakes from seismic data (Filina et al., 2007; Peters et al., 2008; Horgan et al., 2012), although with limited indication of thickness and age, and each lacking seismic analysis of sediment composition. Consequently, the stability
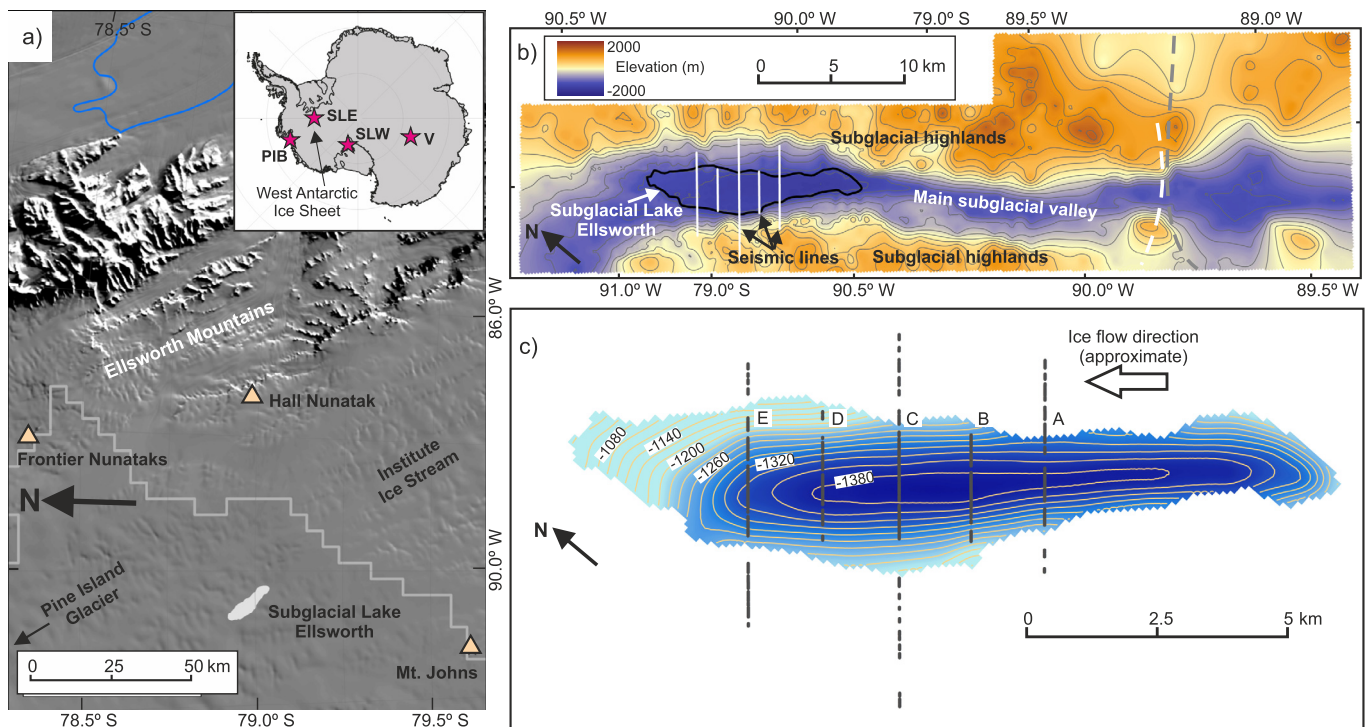
Fig. 1. Location maps for Subglacial Lake Ellsworth. a) Location of Subglacial Lake Ellsworth relative to adjacent geographical features. Grey line is the ice divide, blue line is the Ronne Ice Shelf grounding line and brown triangles are nearby nunataks. Inset shows the lake's location within Antarctica as well as other subglacial lakes mentioned in the text: Subglacial Lake Ellsworth (SLE), Vostok Subglacial Lake (V), Subglacial Lake Whillans (SLW), Pine Island Bay palaeo-subglacial lake (PIB). b) Elevation of the ice base (m above WGS84 ellipsoid). Contour interval is 200 m. For grounded ice this is the subglacial bed topography; for floating ice, this is the lake surface. Black line is the lake outline; white lines show the locations of the seismic reflection lines; grey dashed line is the ice divide; white dashed line is the subglacial hydrological divide. c) Elevation of the lake bed (m above WGS84 ellipsoid) derived from the seismic data. Grey lines (labelled) show the places where bed topography was picked on the seismic lines, both within and outside the lake. The topography presented in b), is effectively the same as that presented previously (e.g. Siegert et al., 2012). The lake bed topography in c) differs from that in previous publications (e.g. Woodward et al., 2010) as it was derived by gridding the actual bed topography data points, rather than by subtracting the gridded water column from the spatially more-extensive lake surface data. This more accurately reflects the limited data on which the lake bed topography map is based. (For interpretation of the colours in the figure(s), the reader is referred to the web version of this article.)

and longevity of subglacial lakes in the ice sheet interior, and hence their significance as repositories of ice sheet history and unique microbial communities, remain un-proven and unquantified.

Subglacial Lake Ellsworth (SLE) lies under the West Antarctic Ice Sheet (Fig. 1). It is approximately 15 km long, 3 km wide, up to 156 m deep, and ice thickness is 2930–3280 m (Woodward et al., 2010). Radar and GPS surveys (Rivera et al., 2006; Vaughan et al., 2007; Woodward et al., 2010; Ross et al., 2011a; Siegert et al., 2012) have mapped the ice surface, thickness and flow; seismic reflection surveys have measured the water depth, giving the lake bathymetry and enabling modelling of water circulation and ice–water interaction (Woodward et al., 2010). One attempt to access SLE has been made but was unsuccessful (Siegert et al., 2014). With reference to limnological and glacial processes, the nature and distribution of sediments in SLE, SLW and Vostok Subglacial Lake have been considered, and conceptual sedimentary environment models proposed (Bentley et al., 2011).

In this paper we analyse seismic reflection strengths and interpret the nature of the lake bed and surrounding subglacial material. We investigate the lake's sedimentary environment, interpret lake bed sediment thickness and speculate on the sediment age. We use our results to develop a new conceptual subglacial lake model, and consider the implications for SLE and for subglacial lakes in general. This provides a new framework on which to interpret future samples, and an established sequence of techniques to understand the environments in, and prepare for direct access to, other Antarctic subglacial lakes.

2. Data and methods

2.1. Seismic data acquisition and processing

Five seismic reflection lines were acquired over SLE in 2007–08 (Fig. 1), three of which extended from the lake onto the adjacent grounded areas. The seismic source was 0.45 kg of high explosive, buried in 30 m-deep holes. The seismic energy from each shot was detected by 48 geophones (40 Hz natural frequency) spaced at 10 m intervals, placed at distances between 30 m and 500 m from the source. Data were digitised and logged at 4 kHz. Shot spacing was 240 m, resulting in single-fold coverage. Data processing included normalisation for shot-to-shot variability (using the energy in the direct wave for each shot), normal moveout correction and migration. A small number of larger shots (1.05 kg) were used to determine attenuation within the ice and to calibrate the bed reflection coefficient over grounded ice. The processed seismic sections are given in Fig. 2.

2.2. Determining the properties of the subglacial material

The strength of the reflection from the lake bed can indicate its acoustic properties and hence, allow interpretation of the likely material there. The energy of the seismic source is not known, so a means of calibrating the reflection strength was needed. For the lake bed, the ice–water interface was used for this calibration. At normal incidence (which is the case for these data), the seismic reflection coefficient, R, of an interface is given by

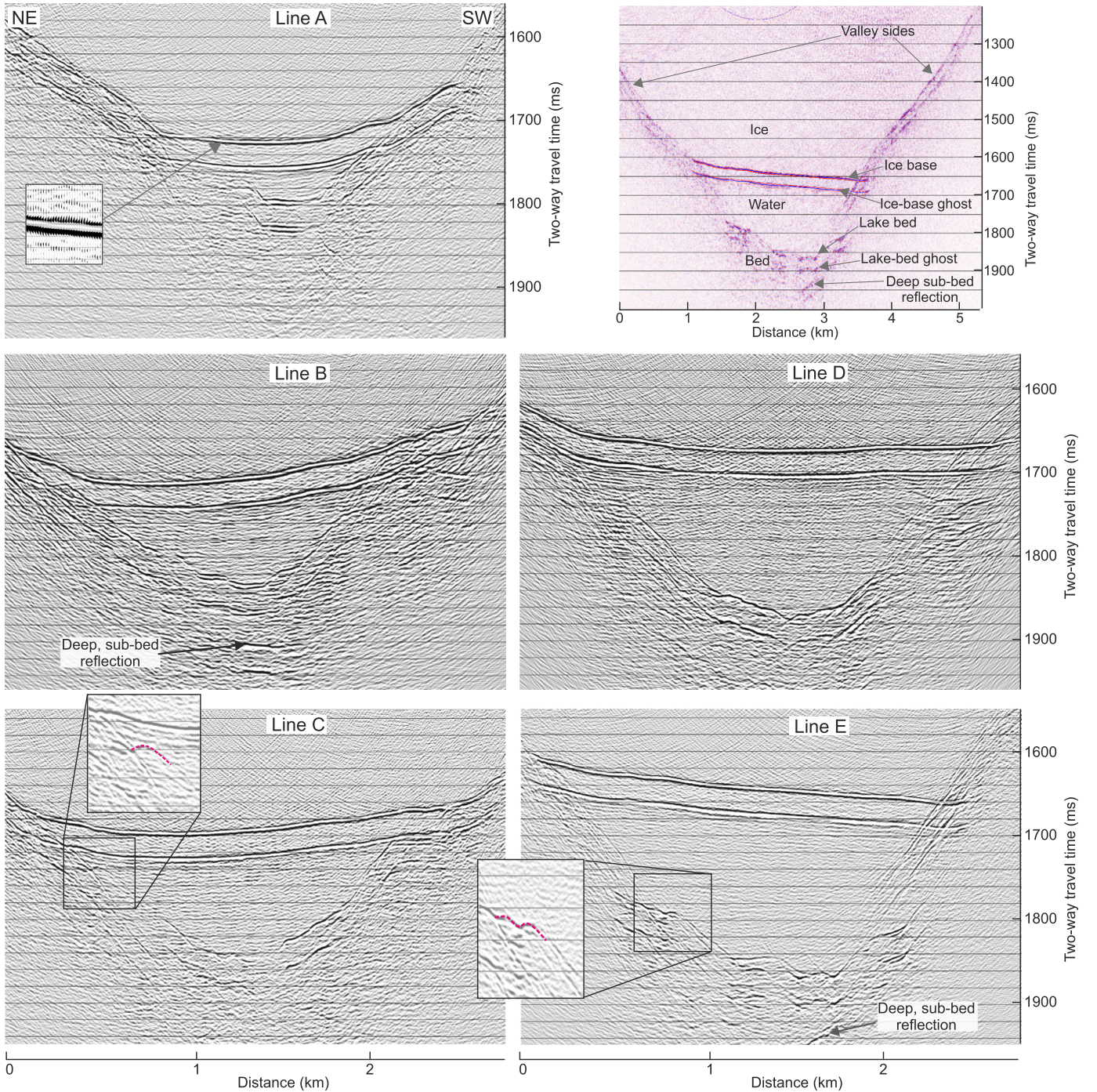$$R_{xy} = (Z_y - Z_x)/(Z_y + Z_x),   \qquad (1)$$

Fig. 2. Processed seismic reflection sections. The parts of the seismic section covering the subglacial lake and nearby grounded ice are shown for each line (A–E); horizontal and vertical scales are the same. Coloured panel (top right) shows the complete section for Line E, with the main reflections identified. On all lines, the reflection from the ice base is clear over the whole lake and much of the adjacent grounded ice. The reflection from the lake bed is more variable. It can be clearly identified over more than half the lines, although it is particularly hard to identify close to the lake shoreline. Source ghosts (artefacts of the acquisition geometry) are clear in many places, occasionally masking the lake-bed reflections, particularly close to the shoreline and where there are indications of deeper, sub-bed arrivals. Insert on Line A shows the detail of the ice–water reflection. In this display the central white band indicates a reversed-polarity reflection, as expected for an ice–water impedance contrast; this central peak was used to calculate the reflection energy. Small sub-sections on lines C and E show sediment mounds (dashed red lines) identified at the lake bed.

where Z_x is the acoustic impedance (product of seismic velocity and density) of the incident medium and Z_y is that of the reflecting medium. SLE contains fresh water (Vaughan et al., 2007), for which density is 1013 kg m^{-3}, and the seismic velocity is 1437 m s^{-1}, giving the acoustic impedance of the lake water, $Z_w = 1.46 \times 10^6$ kg m^{-2} s^{-1}. Assuming seismic velocity and density for ice of 3800 m s^{-1} and 917 kg m^{-3}, respectively, gives the acous-

tic impedance of the ice, $Z_i = 3.5 \times 10^6$ kg m^{-2} s^{-1}, and hence the reflection coefficient of the ice–water interface, $R_1 = -0.41$. The reflection coefficient at the lake bed (R_2) is related to the difference in energy between the ice–water and water–bed reflections by

$$R_2^2 = \frac{E_2 R_1^2}{E_1 (1 - R_1^2)^2} \left(1 + \frac{V_w h_w}{V_i h_i} \right)^2 \qquad (2)$$

E_1 and E_2 are the energies of the ice–water and water–bed reflections, respectively; V and h are the seismic velocity and thickness of the ice (i) and water (w). For each trace where both reflections were clear, the energy of each was given by the sum of the squared amplitude values for the central peak of the wavelet (between the second and third zero-crossings; Fig. 2). The water column thickness was calculated from the difference in travel time between the two reflections and the seismic velocity of the water. Equation (2) then gave the lake-bed reflection coefficient for each trace. Knowing this, and the acoustic impedance of the lake water, rearranging Equation (1) gave the acoustic impedance of the lake bed material.

Outside the lake area, where the ice is aground, another approach was used to determine the properties of the bed material. As with the lake bed, the strength of the ice–water reflection over the lake was used as a reference; corrections were made at each grounded ice location for the difference in ice thickness from that over the lake (Smith, 1997). This required knowledge of attenuation in the ice, a, which was determined using the ice–water reflection from a number of shots that recorded both the primary reflection and its first multiple (energy which reflects off the ice sheet surface and then off the lake a second time). The reduction in energy between the primary (E_1) and multiple (E_1') arrivals is related to ice thickness, attenuation and ice–water reflection coefficient by

$$\frac{E_1}{E_1'} = \frac{4}{R_1^2} e^{4ah} \tag{3}$$

This analysis gave a value for a of 0.28×10^{-3} m^{-1} (SD = $\pm 0.07 \times 10^{-3}$ m^{-1}; $n = 245$), similar to that derived elsewhere in West Antarctica (e.g. Bentley and Kohnen, 1976) and assumed to be a typical value for this region of the ice sheet (e.g. Smith et al., 2013).

Table 1

Acoustic impedance results for the lake bed and surrounding subglacial material from the seismic reflection analysis, and the interpreted lake bed and subglacial materials. Mean values, with Standard Deviation, are given for each interpreted region. See Fig. 3 for individual data points and distances over which mean values were calculated.

Acoustic impedance		Interpretation
Mean, with Standard Deviation (10^6 kg m^{-2} s^{-1})	Number of points	
1.5 [0.3]	110	Lake water
2.1 [0.4]	582	Lake bed: subaqueous, soft wet sediments
4.7 [0.6]	160	Subglacial, soft wet sediments
6.5 [1.7]	358	Hard bed: wet sediments
8.7 [3.7]	33	Hardest bed: rock or frozen sediment

3. Results

We used the reflections from the three main interfaces: ice–water, water–bed and ice–bed (Fig. 2), to enable interpretation of the lake bed and surrounding subglacial material. The acoustic impedance of the bed, determined from the seismic sections is shown in Fig. 3 (see also Table 1) and summarised in Fig. 4, including likely bed materials interpreted from the data shown in Fig. 5a. Additional, analogous data from sub-aqueous environments (Fig. 5b) allowed further interpretation of the composition of the lake bed material.

3.1. Soft wet sediments at the lake bed

The material at the lake bed has low acoustic impedance values (2.0–2.5×10^6 kg m^{-2} s^{-1}; mean 2.1×10^6 kg m^{-2} s^{-1}), similar to those found in ocean floor sediments, freshwater lakes and other subaerial aquatic locations (e.g. Hamilton, 1971; Richardson and Briggs, 1993; Smith, 1997 and references therein; Vardy, 2015).
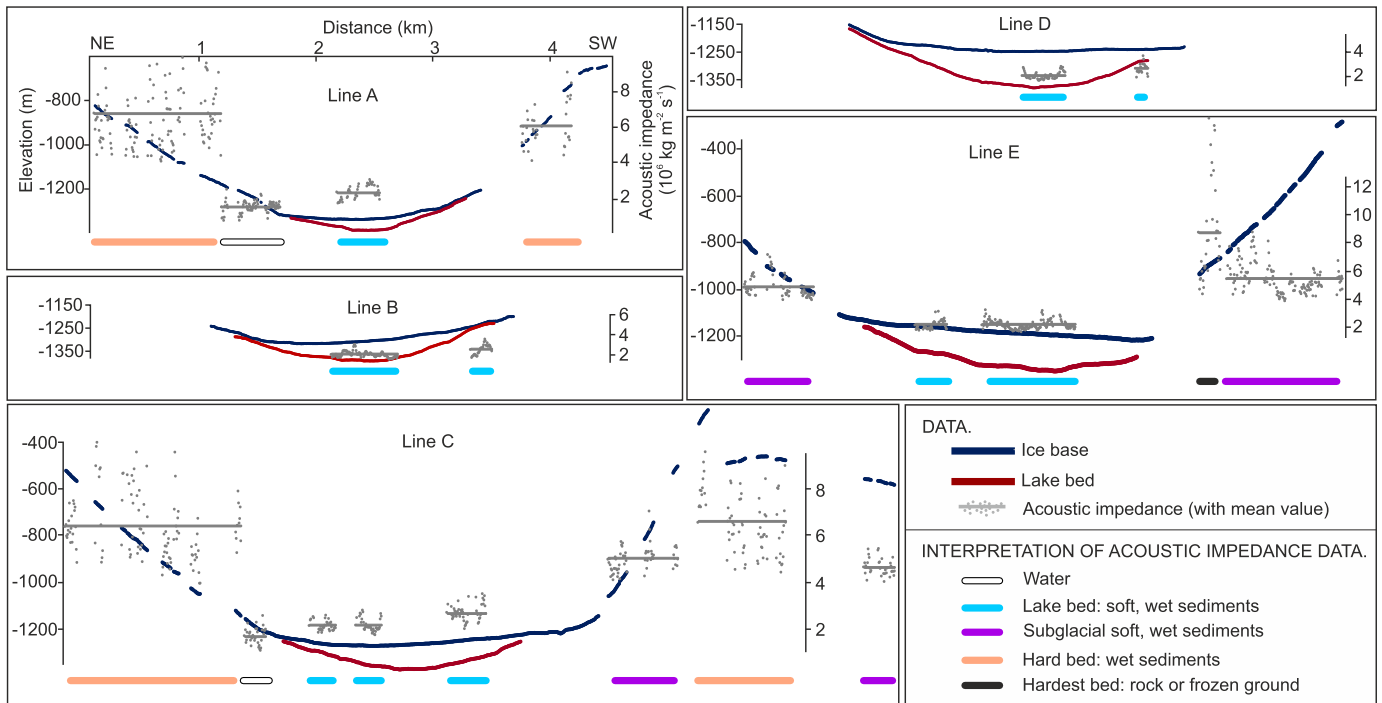

Fig. 3. Lake bed and subglacial material interpreted from the seismic reflection strengths. Orientation and horizontal scales are the same for each line. Left-hand vertical axes are elevation above WGS84 datum; right-hand vertical axes are the acoustic impedance determined from reflection strengths. Ice base and lake bed in the vicinity of the lake are continuous, interpolated values (Woodward et al., 2010); outside the lake the actual picked points of the ice bed are shown. Gaps occur where clear reflections from the bed could not be identified, or were insufficiently defined for their reflection strength to be measured. Mean values over each region of the bed, and an indication of the variance in the data, are shown in Table 1.
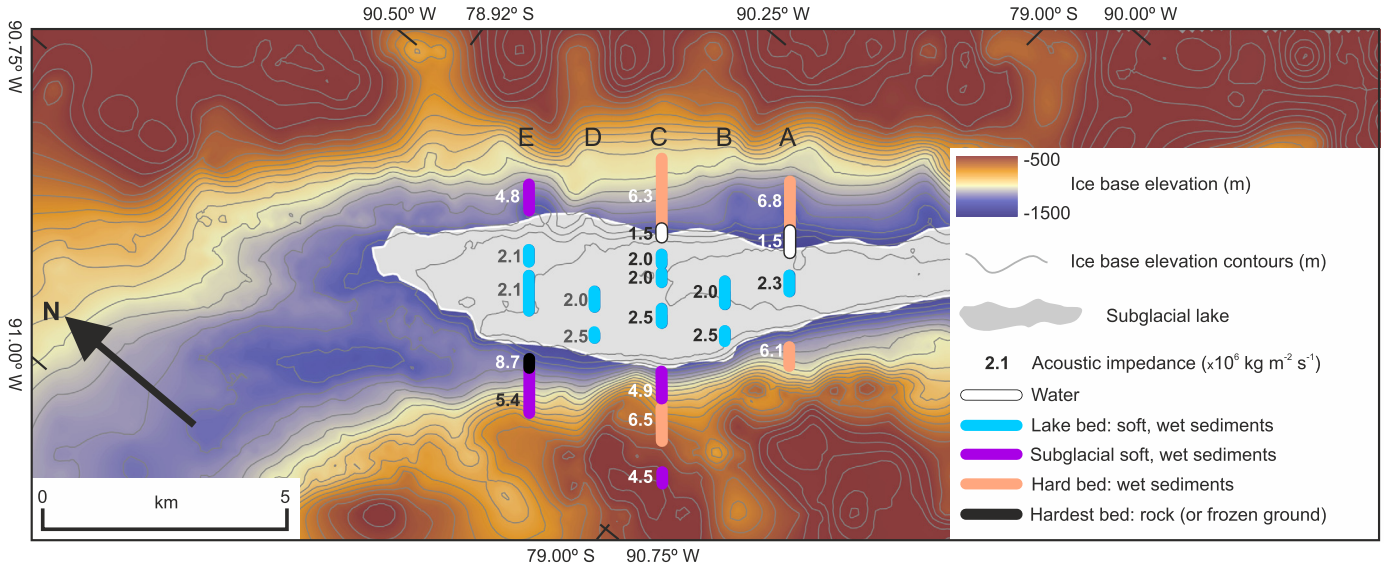
Fig. 4. Distribution of bed materials beneath and surrounding SLE interpreted from the seismic reflection strengths. Elevation is relative to WGS84, contour interval is 100 m. Numbers are mean acoustic impedance ($\times 10^6$ kg m^{-2} s^{-1}) for each region. Coloured lines over the lake show the different interpreted lake bed and subglacial materials.
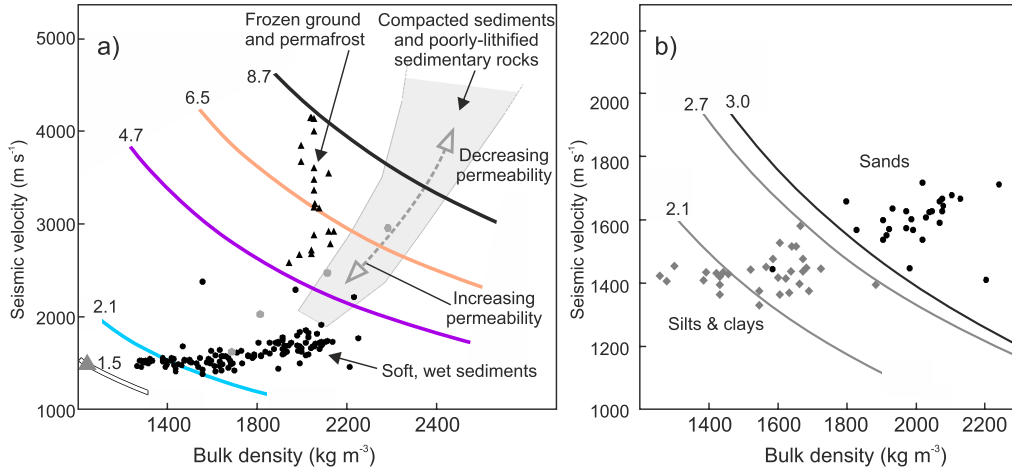

Fig. 5. Acoustic impedance of sediments. a) Measured values of seismic velocity and density for a range of sedimentary materials (LeBlanc et al., 1992; Richardson and Briggs, 1993; Smith, 1997 and references therein; Smith et al., 2002 and references therein), including different sediments, poorly-lithified sedimentary rocks, permafrost, frozen ground and water, compared with results from SLE (coloured curves). Coloured curves show lines of constant acoustic impedance corresponding to the mean values determined from the seismic reflection strengths; colours correspond to those in Fig. 4. Black circles are soft, high porosity, wet sediments. Black triangles are frozen ground and permafrost. Grey circles are poorly lithified, Cenozoic sedimentary rocks from McMurdo Sound, Antarctica. Grey triangle is water. Grey area is the lower end of typical values for compacted sediments and soft, poorly lithified sedimentary rocks (e.g. sandstones and shales). b) Seismic velocity and density for soft, wet subaqueous sediments in a) for which the sediment composition is also known. Black circles are sands and sandy sediments, grey diamonds are silts and clays. Curves show selected lines of constant acoustic impedance (2.7×10^6 kg^{-2} s^{-1} and 3.0×10^6 kg^{-2} s^{-1}) that separate sandy and fine-grained sediments, for comparison with the results from the bed of SLE (2.1×10^6 kg^{-2} s^{-1}).

Sediments with acoustic impedance values in the range 2.0–3.0×10^6 kg m^{-2} s^{-1} typically have very high porosities (40–80%); seismic velocity is dominated by that of the water, and is normally ~ 1500 m s^{-1} at the sediment surface, increasing only gradually with depth (still typically <1600 m s^{-1} at 100 m depth; Hamilton, 1979). Most of the variation in acoustic impedance in such sediments comes from bulk density differences resulting from variations in porosity and, to a lesser extent sediment grain type (e.g. quartz or clay minerals).

Most sandy, sea floor and subaerial lake sediments (coarse, medium and fine sands, silty-sands) have acoustic impedance values of $\geq 3 \times 10^6$ kg m^{-2} s^{-1} (Fig. 5b); in contrast, silt- and clay-dominated sediments have impedance values of $\leq 2.7 \times 10^6$ kg m^{-2} s^{-1}. All measured regions of the lake bed show values of $\leq 2.5 \times 10^6$ kg m^{-2} s^{-1}, indicating that these sediments are all fine grained, silts or clays (particle diameter less than ~ 60 µm)

with no more than a minor proportion of larger grain sizes. This is consistent with the homogeneous, fine-grained sediment recovered from a palaeo-subglacial lake in Pine Island Bay (Kuhn et al., 2017; porosity 77%, clay–silt–sand 80–18–1.6%).

The low acoustic impedance values therefore indicate that the lake bed material in the area covered by the seismic surveys is high porosity, fine grained sediment. Establishing and maintaining such a low impedance requires slow sedimentation in a low energy aquatic environment over a long period of time; other conditions, or physical disturbances, would lead to a reduced porosity and higher impedance value. Low sedimentation implies low sediment flux reaching this region of the lake bed; combined with the interpreted low-energy environment, this presumably also implies a relatively low water flux entering the lake. Hence we interpret the lake bed, in this region at least, as fine-grained sediments de-

posited in a long-term, low energy environment with low water- and sediment-fluxes.

3.2. Subglacial soft wet sediments outside the lake

Surrounding the lake in the downstream- and some of the mid-region covered by the seismic surveys, the ice sheet bed has acoustic impedance values of $\sim$4.5–5.5 $\times 10^6$ kg m^{-2} s^{-1} (mean 4.7 $\times 10^6$ kg m^{-2} s^{-1}). These values are typical of water-saturated, permeable sediments (Fig. 5a) with lower porosity than those on the lake floor. The upper end of this range could include saturated sediments in which some of the water in the matrix is frozen, but overall the sediment remains permeable (Smith et al., 2002). These areas show that the ice surrounding the lake is not frozen to the bed and that the ice-bed interface does not form an impermeable seal to the lake water. Depending on hydrological potential gradients, water can move between the lake and its surroundings, either at the shoreline or by groundwater flow through adjacent permeable sediments.

3.3. Harder subglacial sediments

The acoustic impedance of the bed surrounding the lake in the upstream region of the seismic surveys lies in the range $\sim$6–7 $\times 10^6$ kg m^{-2} s^{-1} (mean 6.5 $\times 10^6$ kg m^{-2} s^{-1}). These values are typical of well-compacted sediments, very poorly lithified sedimentary rocks, or frozen ground. Fully frozen ground is likely to be impermeable, whereas weak sedimentary rocks will be unfrozen with a thawed interface. Whilst the acoustic impedance data alone cannot distinguish between these alternatives, the close proximity to soft, water-saturated sediments and to the lake itself, makes it most likely that this subglacial bed is unfrozen sedimentary material. Hence, given suitable hydrologic potential gradients, water can move across the shoreline in these areas.

3.4. Hardest subglacial material

The ice bed adjacent to the lake's NW shoreline on Line E has a mean acoustic impedance of 8.7 $\times 10^6$ kg m^{-2} s^{-1} (range $\sim$4–17 $\times 10^6$ kg m^{-2} s^{-1}) significantly higher than anywhere else covered by the seismic surveys. This value is typical of bedrock, or possibly substantially-frozen sediment. As the permeability of either is likely to be low, this may limit or even preclude the movement of water into or out of the lake at this location. This area appears to be part of a substantial bed ridge mapped by radar surveys and interpreted as a sill formed of impermeable material that acts as a dam impounding the downstream end of SLE, restricting water outflow (Ross et al., 2011b). The limited permeability indicated by the seismic data, further supports this interpretation.

3.5. Lake water

Two areas (on lines A and C; Fig. 4), previously identified as grounded ice on the basis of radar reflection characteristics (Woodward et al., 2010; Ross et al., 2011a, 2011b), show acoustic impedance values expected for water (1.5 $\times 10^6$ kg m^{-2} s^{-1}). These locations also lie landward of significant changes in the slope of the ice base, where the ice is supported well above the level for floatation (Vaughan et al., 2007). In these areas we see no clear seismic reflection from the lake bed; however, we can interpret the minimum water depth using the ice–water reflection characteristics. The duration of the reflection wavelet in the seismic data is $\sim$8 ms, equivalent to a water depth of 6 m. Any reflection from an interface within that distance (from the lake bed, for example) would distort the wavelet shape, which we do not see. Hence, to give the clear reflection and the acoustic impedance results we

found, there must be a layer of free water at least 6 m thick at these two locations.

4. Interpretation

4.1. The sedimentary environment in Subglacial Lake Ellsworth

Soft, high porosity, fine grained sediments cover the whole of the lake bed area in the region of the seismic surveys. Their presence and widespread distribution are evidence for a low energy environment with little disturbance, low water- and sediment-fluxes, and slow, steady sedimentation persisting over an extended period of time. This is identical to the conditions interpreted for a palaeo-subglacial lake in Pine Island Bay (Kuhn et al., 2017).

The presence of permeable sediments around the lake means that SLE is not strictly a closed hydrological system. In a closed system, water can only enter and leave the lake by melting and freezing at the ice–water interface; in an open system, water can also enter or leave as drainage at the ice-bed interface or as groundwater flow through the bed. The nature of the hydrological system is relevant for the origins of the lake-bed sediment, likely sources of which are (Bentley et al., 2011): melt-out of dust originally deposited at the ice sheet surface, melt-out of sediment entrained into basal ice from the surrounding bed, chemical sedimentation, advection of subglacial sediment, and transport of sediment in subglacial water. The latter two sources can only occur in an open system, whereas the rest could occur in one that is either open or closed. Chemical sedimentation, although interesting mineralogically and biogeochemically, is probably volumetrically minor (Bentley et al., 2011). Sediment melt-out will depend on debris concentrations within the ice and the rate and area of basal melting. Advection of subglacial sediment requires an upstream source of mobile material and will depend on ice flow rates. Transport by subglacial water requires an active hydrological network and a sediment source, but if these are sufficient, this could be the dominant mechanism for delivering sediment into subglacial lakes (Bentley et al., 2011). Our results add little new to estimates from chemical sedimentation and entrained material (both surface- and basally-derived); however they do allow significant new considerations of water transport and (to a lesser extent) basal advection, and henceforth we will concentrate mainly on those mechanisms.

4.2. Subglacial sediment and water sources

The material present at the lake bed will be related to the sources and fluxes of sediment, and of the water and ice that transport it. Likely source regions are the main valley SE of the lake and the adjacent subglacial highlands. SLE lies close ($\sim$20 km) to the hydrological divide (Fig. 1) and its catchment is small ($\sim$200–450 km^2; Bentley et al., 2011; Vaughan et al., 2007). Estimates of geothermal heat and hence, basal melt, although poorly constrained, are low (e.g. Pattyn, 2010), suggesting that water fluxes into the lake are also low. Hydrological potential (Siegert et al., 2012), assuming a thawed bed and zero effective pressure, shows that the only subglacial water flow-path of significant length entering SLE is that from the SE ($\sim$20 km long) along the main valley. Flow-paths from the adjacent highlands are short ($<$5 km), drain only small areas and, as the ice is much thinner than in the main valley, may even be frozen in places (Pattyn, 2010). Assuming that water flux at a point is related to the cumulative upstream path length and area drained, the upstream valley is likely to be the main source of any water-transported sediment into SLE. Subglacial sediment can also be transported by ice movement. Ice flow is predominantly along the main valley so material advected into the lake this way is also likely to be mainly from that source.

The pre-glacial geology will be the main factor determining sediment composition. In the region of the lake, the main valley is sufficiently deep (-1393 m in the lake bed) to have remained well below sea level prior to the last glaciation, when it will have had an open pathway to the ocean (Vaughan et al., 2007), with consequent marine sedimentation. Further upstream, the valley floor rises, reaching a high point of around -800 m (Ross et al., 2011b). This may have been close to sea level when last ice free (Ross et al., 2014) and subglacial sediments there could also include those typical of high-latitude fjord environments (Syvitski and Shaw, 1995). Hence, lake bed sediments derived from the main valley could comprise a range of materials including fine-grained marine, shallow marine and coarser terrestrially-derived ones. Evidence for bedrock in the region includes the nearby quartzite outcrop of Mt Johns (Anderson, 1960), similar to rocks exposed in the Ellsworth Mountains. Potential field data suggest that similar bedrock underlies the ice sheet around SLE (Jankowski and Drewry, 1981). Hence, in contrast to sediments derived from the main valley, erosion of these rocks, could result in relatively coarse, sandy material.

All the lake bed sediments in the area of the seismic surveys are fine silts and clays, consistent with a source further up the main valley, with little coarse-grained sediment from the adjacent highlands. None of the acoustic impedance data reach the lake shoreline, the closest being $\sim$300 m (Line B, SW end). Although we expect very little coarse sediment from the lake sides, any that is delivered would be deposited within a short distance of entering the lake and remain undetected by our seismic results. In addition, Line A is >5 km from the upstream shore and, as the shallower parts of the upstream valley could contain terrestrially-derived sediment, coarser deposits remain possible at the upstream end of the lake.

4.3. Processes within the upstream part of the lake

The seismic data cover only a small proportion of the bed of SLE, comprising just five, 2-D seismic lines. This lack of coverage is most significant for the upstream part; the first seismic line (Line A) is more than one third of the way down the lake. The presence of a grounding-line fan or morainal bank at the upstream shoreline has been hypothesised (Bentley et al., 2011), formed from coarse sediment entering the lake there either by advection or entrained in subglacial water. Subglacial water will be at its pressure-melting point and hence neutrally buoyant on entering the lake (Thoma et al., 2011). Sediment load however, will increase the density leading to an underflow, where incoming water flows across the lake bed, encouraging rapid deposition of any coarse material. The seismic data confirm that such coarse deposits, if they exist, do not extend as far as the location of the first seismic line.

The other possible source of coarse material at the lake bed is melt-out of subglacially entrained material. While we have no direct evidence to quantify this, the lower 5 m of the Byrd ice core contained up to 7% sediment by volume (Gow et al., 1979). Assuming an ice flow speed over the lake (Woodward et al., 2010; Ross et al., 2011a, 2011b) of 5 m a^{-1} and basal melting over the upstream part (Thoma et al., 2011) of 5 cm a^{-1}, a similar thickness of sediment-laden ice would have all melted within 500 m of the upstream grounding line.

4.4. Processes within the middle part of the lake

The bed in the middle part of the lake has ubiquitous soft, fine-grained sediments deposited in prolonged low-energy hydrological conditions with low water- and sediment-fluxes. These sediments will be rain-out from material suspended in the water column, likely sources of which are subglacial water and melting of basal ice releasing its surface-derived dust component. Estimates of dust sedimentation rates are extremely low, e.g. 10 cm Ma^{-1} in East Antarctica (Bentley et al., 2011). Even allowing for possible higher rates in West Antarctica and for the nearby Ellsworth Mountains as an enhanced dust source, this contribution for SLE is likely to be volumetrically small and will only occur where the ice base is melting (Woodward et al., 2010). Any coarse sediment in subglacial water will be deposited soon after it enters the lake, leaving fine particles in suspension. Hence, we believe that rain-out of material carried into the lake by subglacial water is the source of most of the sediment accumulated at the bed of SLE.

4.5. Processes within the downstream part of the lake

Filling and draining events are known to occur in some subglacial lakes (e.g. Smith et al., 2009), sometimes flowing between individual lakes (Wingham et al., 2006). Filling and drainage rates probably cover a wide range of values, and observations are currently too sparse to indicate if typical rates exist; however, those limited observations have shown filling events occurring over a number of years and drainage events lasting a few days, often involving high water fluxes. These are major, high-energy turbulent events capable of disrupting soft sediment sequences. Any physical disruption in the lake is likely to result in an increased acoustic impedance; hence, the fact that there is widespread soft, fine-grained sediment at the bed of SLE, rather than bedrock or other well-lithified material, is evidence against discrete, high energy events, such as discharge of large volumes of water into or out of the lake. It has been hypothesised that sediments at the downstream grounding line, where the lake is impounded by the impermeable bedrock ridge, could show truncation of the sedimentary sequence resulting from erosion during rapid drainage events (Bentley et al., 2011). Our conclusion that such events are unlikely to have occurred suggests truncation of beds by this mechanism is unlikely.

4.6. Water circulation and fluctuation

The identification of an open hydrological system for SLE has implications for models of water circulation and ice–water interaction which assume it is closed (Woodward et al., 2010; Thoma et al., 2011). The model outputs are unlikely to be quantitatively correct, although geothermal heat and the steep ice–water interface ($>2°$) mean that the modelled processes (circulation and ice–water interaction) do probably occur, but at unknown rates.

Filling and draining of subglacial lakes are normally identified by rising and falling of the ice sheet surface. There is no evidence that this is happening in SLE; if it is, changes are too slow to be detected. Slow drainage might leave little evidence in the seismic data; recently-grounded lake bed sediments would quickly increase compaction due to the overlying ice, reducing porosity and increasing acoustic impedance, making them indistinguishable, seismically, from the current soft, wet subglacial sediments. Lake filling, however, would leave areas of recent un-grounding, still over-compacted from the weight of overlying ice; this would be indicated by higher acoustic impedance values. Of all the places where we have determined the lake bed acoustic impedance, the SW end of Line B has the thinnest water column ($\sim$6.5 m). The acoustic impedance shows that the bed material there is the same as elsewhere across the lake, with no indication of over-compaction. Hence, whilst we cannot say whether or not SLE is slowly draining, we can conclude that it has not undergone significant net filling, for a long period of time.
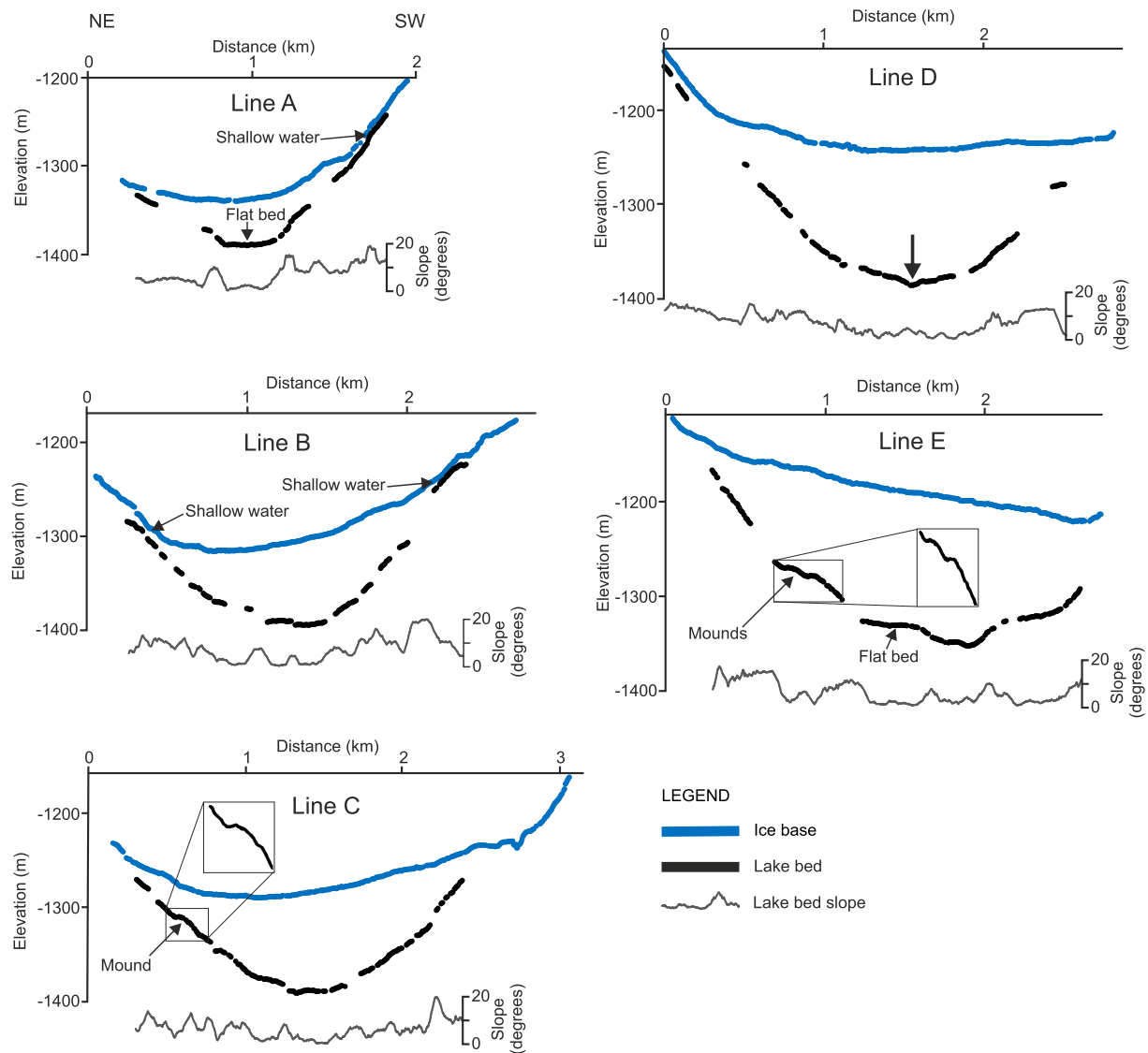
Fig. 6. Ice base and lake bed topography along the SLE seismic lines. The graphs show actual picked data points, rather than continuous, interpolated profiles. Elevation (vertical scales) is relative to WGS84. The slope of the lake bed for each line is shown, derived from smoothed (50 moving average) interpolated data (Fig. 3). Vertical arrow in Line D is the previously-proposed lake bed sampling location (Woodward et al., 2010).

4.7. Lake bed topography and mass movement

At a few locations (Fig. 2, Fig. 6) there are discrete mounds at the lake bed, typically 10 m high and 100 m wide, and where acoustic impedance is not significantly different to adjacent areas or to other locations on the lake bed. These features do not correlate between adjacent seismic lines indicating they are discrete individual features, rather than more-extensive, linear forms. There are two likely explanations for these mounds: mass movement of the lake bed sediments, or a pre-existing surface, formed before the lake, that has since been buried by lake sedimentation, yet retains the earlier topography.

We cannot distinguish unequivocally between these two possibilities, but we believe that mass movement is the more likely explanation. The shape of the mounds, in which the trailing, up-hill sides have a reverse slope (i.e. opposite to the overall slope gradient), is typical of sediment mounds caused by mass movement in fjord environments, where large slump events create characteristic deposits; these are seen both in recent slides and in ones buried by subsequent sedimentation (Syvitski and Hein, 1991; Bellwald et al., 2016). The mean slope angle of the lake bed on

the seismic lines (excluding the areas close to the shoreline where the bed could not be identified) is 7.5°, with a typical range of 5–20° (data interpolated and smoothed with a 50 m running mean; Fig. 6). These are moderate slope angles, at which mass movement features have been reported in other lakes (e.g. Kremer et al., 2015), although they are infrequent (Schnellmann et al., 2006). Hence, for subglacial lakes like SLE, sediment mass movement structures should probably be expected.

In a subglacial lake, sliding could be induced by earthquakes, water seepage through the bed or spontaneous slope failure from long-term accumulation of sediment (Kremer et al., 2015; Schnellmann et al., 2006). All of these are likely to be infrequent. Spontaneous triggers are rare, or absent when sedimentation rates are low (e.g. Bellwald et al., 2016). Groundwater seeping through the lake bed is probably low, otherwise changes in surface elevation, or variations in acoustic impedance would be expected. Finally, except for microearthquakes associated with fast ice flow, Antarctica has no greater levels of seismicity than elsewhere on Earth (e.g. Kanao, 2014), and earthquake-induced triggers are expected to be infrequent. Hence, any of these trigger mechanisms could cause the mounds at the bed of SLE, but they are likely

Table 2
Linear sedimentation rates (LSR) from possible analogues for subglacial lake sedimentary environments and corresponding lengths of time to accumulate 6 m of sediment in SLE. Pine Island Bay LSR was calculated assuming at least 343 cm of subglacial lake sediment at core site PS69/288 were deposited prior to 11.3 ka, from either the end of the last interglacial (71 ka) or the beginning of the last glacial period (29 ka). Sediment accumulation times given in grey are older than the Antarctic Ice Sheet (~34 Ma) and hence, considered unlikely for SLE.

Location	Environment	LSR (mm ka^{-1})	Time required for 6 m of sediment accumulation
Great Slave Lake[a]	Former subglacial lake, Laurentide Ice Sheet	2000	3 ka
Pine Island Bay[b]	Palaeo-subglacial lake at periphery of present-day Antarctic Ice Sheet	>57–194	30–105 ka
Lake Hodgson[c]	Subglacial lake at periphery of present-day Antarctic Ice Sheet	41	147 ka
Larsen B Ice Shelf[d]	Sub-ice shelf Holocene sedimentation, Antarctica	30	200 ka
Ocean floor	Deep-sea argillaceous sedimentary environment	5	1.2 Ma
Abyssal plain	Abyssal plain, deep ocean floor	1	6 Ma
Progress Lake[e]	Subglacial lake at periphery of present-day Antarctic Ice Sheet	0.11	54 Ma

[a] Christoffersen et al. (2008).
[b] Kuhn et al. (2017).
[c] Hodgson et al. (2009).
[d] Domack et al. (2005).
[e] Hodgson et al. (2006).

to be rare events. Any physical disturbance, including sliding and re-settling, is likely to increase the acoustic impedance of the sediment; as the impedance values over the mounds are the same as elsewhere on the lake bed, we conclude they are most likely to be old features, buried by subsequent sedimentation.

5. Sediment thickness and age

5.1. Sediment thickness

The seismic reflection wavelet allows us to estimate the minimum thickness of soft, fine grained sediments at the lake bed. The wavelet duration (8 ms) is equivalent to 6 m in the sediments. As with the water layer on lines A and C, any arrival from a reflecting interface within that depth would distort the wavelet shape, which we do not see. Hence, 6 m is the minimum sediment thickness we interpret at the bed of SLE. Clear reflections from deeper within the bed only occur at two locations (Fig. 2). These arrive at two-way travel times of 60–100 ms after the lake bed. If they represent the bottom of the sediment sequence, this would imply a thickness of 50–80 m, with a mean value of ~60 m (assuming a seismic velocity of 1.5 km s^{-1}; Hamilton, 1979).

5.2. Sediment age

Sedimentation rates are unknown for present day subglacial lakes in the ice sheet interior so we can only speculate on the age of the sediments in SLE. Linear sedimentation rates (LSR) from a number of possible analogues are given in Table 2, ranging from 2 mm a^{-1} (Great Slave Lake; Christoffersen et al., 2008) to 0.1 mm ka^{-1} (Progress Lake; Hodgson et al., 2006). For each LSR, Table 2 shows the length of time required for 6 m of sediment accumulation. The low acoustic impedance values in SLE indicate slow sedimentation rates in a low-energy environment, persisting over a long period of time, suggesting that the younger ages are less likely than the older ones. This implies that the upper 6 m of sediment at the bed of SLE represents a period of at least 150–200 ka, and possibly >1 Ma, consistent with evidence this area remained ice covered through the last interglacial (DeConto

and Pollard, 2016) and much further back through the Pleistocene (Hein et al., 2016) and Pliocene (Pollard and DeConto, 2009).

Proxy evidence suggests the West Antarctic Ice Sheet (WAIS) formed in the late- or mid-Miocene (~15 Ma; e.g. Haywood et al., 2009), or perhaps as early as the onset of Antarctic glaciation at ~34 Ma (Miller et al., 2005). Throughout its history, highland parts of WAIS, including the Ellsworth–Whitmore mountains area, are believed to have persisted through ice sheet retreat during climate warm periods. If the significant interface identified ~60 m below the lake bed marks the base of subglacial lake sediments, the possible age ranges of this sequence would be ~10 times those indicated in Table 2. Hence we speculate that this deep interface could be a transition from pre-glacial to subglacial conditions, possibly in the earlier part of Antarctic glaciation between 15–34 Ma. However, as SLE lies close to the edge of the Ellsworth highlands, even if ice cover was un-interrupted, subglacial lake conditions were not necessarily continuous over this time period.

6. A conceptual model for the sedimentary environment in Subglacial Lake Ellsworth

In the absence of comprehensive observations and data, conceptual models of subglacial lake sedimentary environments provide templates for what the physical processes in a particular system might be, which can be tested and refined as results and direct measurements become available.

In Fig. 7 we present a new conceptual model of the sedimentary environment in SLE, derived by comparing our seismic results with an earlier model (Bentley et al., 2011). Some parts of the original model are supported by our results, some have been updated and others remain untested and hence, un-changed.

The main features of the revised model include the following. A transverse section in addition to a longitudinal one, with a lake surface and bed that more closely reflect the true SLE topography, including a significant step at the downstream end where the lake is impounded by the impermeable bed ridge. Input of water and sediment occurs mainly at the upstream grounding line; even there, fluxes of both are low. Input of water and sediment from the valley sides are very low. Sediment accumulation across
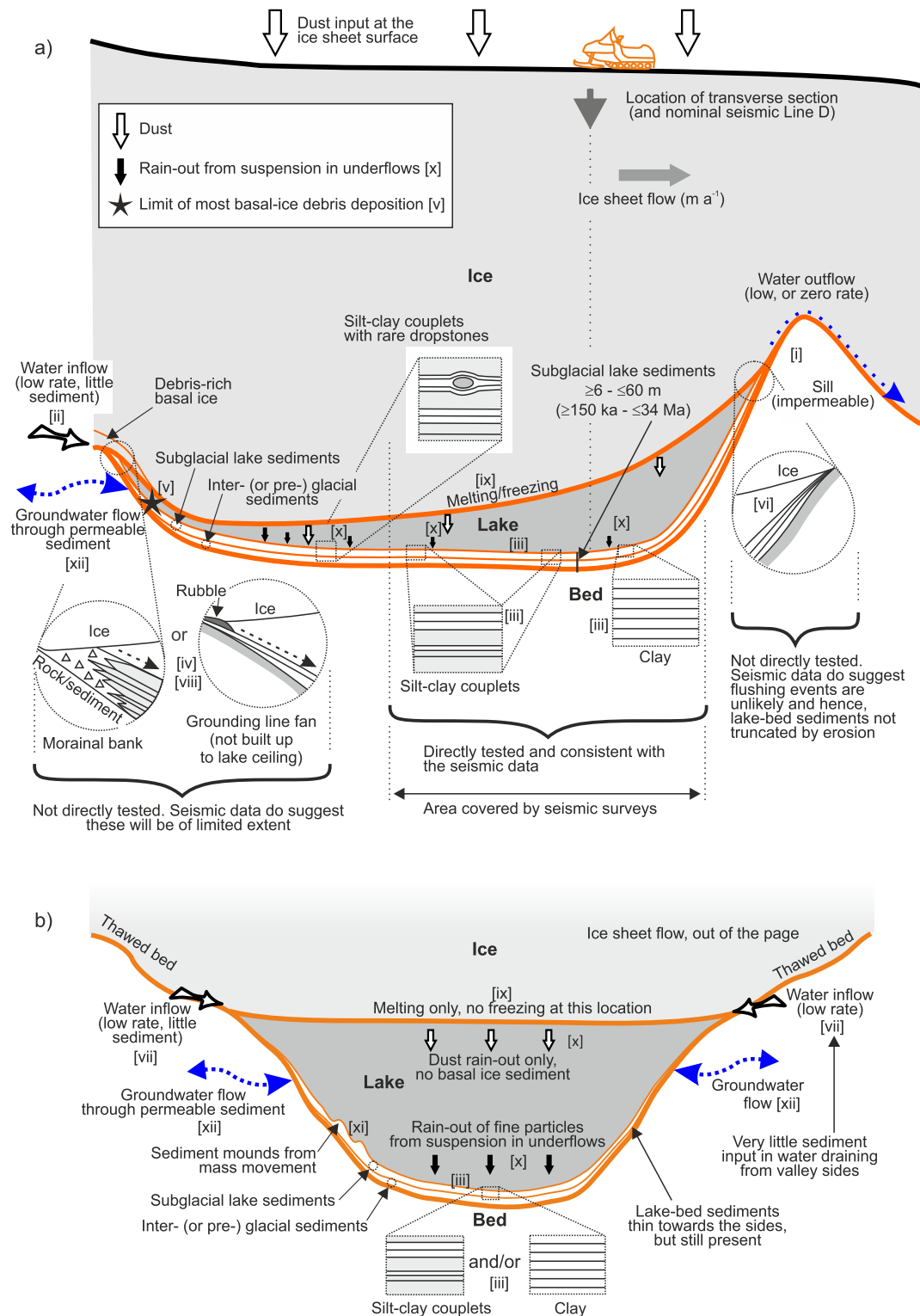
Fig. 7. Conceptual model of the sedimentary environment in SLE. The main features include: a lake surface and bed that more closely reflect the SLE topography [i]; a transverse section in addition to a longitudinal one; input of water and sediment (both of low flux) occurs mainly at the upstream grounding line [ii]; clay- or silt-dominated sediments across the entire lake floor after a short distance beyond the upstream grounding line [iii]; coarse deposits are restricted to within a short distance from the upstream grounding line [iv]; sediment in the basal ice is all deposited within a short distance of the upstream grounding line [v]; high-energy discharge or flushing events at the downstream end are unlikely and the sedimentary sequence there is unlikely to have been disturbed and truncated [vi]; input of water and sediment from the valley sides is very low [vii]; a moraine bank or grounding line fans remain possible at the upstream end [viii], but are less likely at the valley sides; melting of the basal ice occurs over the upstream two-thirds of the lake (Thoma et al., 2011), possibly switching to basal freezing beyond that [ix]; sediment accumulation across the lake floor is dominated by slow rain-out from suspension in underflows with aeolian dust from melting of the ice base [x]; sedimentation rates are very low; mounds at the bed indicate mass movement of sediments that occur infrequently and are old enough to have been buried by subsequent lake sedimentation [xi]; an open hydrological system [ii, ix, xii]. a) Longitudinal section between the upstream (left) and downstream (right) ends of the lake (revised from Bentley et al., 2011). b) Transverse section, nominally at the location of seismic Line D.

the lake floor is dominated by slow rain-out of particles from suspension in underflows. Sedimentation rates are very low. Clay, or silt–clay dominated sediments cover the entire lake floor after a short distance beyond the upstream grounding line. Coarse deposits advected or washed into the lake, and any sediment in the basal ice, are all restricted to within a short distance of the upstream grounding line. A moraine bank or grounding line fans remain possible at the upstream end, but are less likely at the valley sides. Melting of the basal ice occurs over the upstream two-thirds of the lake, switching to basal freezing beyond that. Discharge or flushing events at the downstream end are unlikely; more-steady conditions are expected. Hence, the sedimentary sequence there is unlikely to have been disturbed and truncated by high-energy discharge events. Mounds at the bed indicate mass movement of sediments. These occur infrequently and are old enough to have been buried by subsequent lake sedimentation.

More geophysical data would enable further testing and updating of this model. Direct access and sample retrieval would begin to confirm to what degree the model, and modelled predictions of water circulation and ice base mass balance (Woodward et al., 2010; Thoma et al., 2011), are quantitatively correct.

7. Conclusions

Our results have wide implications for understanding subglacial lakes and their sedimentary records. Structured, fine-grained sediment sequences, resulting from prolonged slow deposition, hold unique records of ice sheet history and microbial communities at these interior ice sheet locations. At SLE, for example, any evidence for sequence disruption could indicate periods of major WAIS retreat. Our analysis of the seismic data, combined with existing results, leads to the following conclusions for SLE.

7.1. SLE sediments

The lake bed comprises soft, wet sediments; there is no evidence for harder material that would be likely had the lake experienced high-energy flushing or drainage events. These lake bed sediments are high porosity, fine-grained silts or clays, not coarser sandy material. They cover the whole lake floor, except perhaps in the upstream part. Sediment mass movement events have occurred, but they are rare and have been buried by subsequent sedimentation. The minimum sediment thickness is 6 m. We speculate that this represents at least 150 ka of accumulation and perhaps a much longer period of time (e.g. >1 Ma). This sequence should in some way reflect subglacial conditions through ice sheet changes back into the last glacial period or earlier. There is some evidence that the sediment could be at least 60 m thick. If this sequence is formed solely of subglacial lake material, it could perhaps represent a period of time as far back as the onset of Antarctic glaciation.

7.2. SLE water

SLE is an open system – water can enter or leave the lake across the shoreline, through a permeable sediment bed, and by ice–water interface processes. As it is an open system, water circulation models which have assumed it to be closed are unlikely to be quantitatively correct. Rates of water and sediment influx to the lake are low. There has been no recent significant net filling of the lake; we cannot discount drainage, provided it is sufficiently slow and low-energy to avoid disturbing the lake bed sediments.

7.3. SLE summary

The seismic data indicate that SLE is a long-lived, low energy environment and that any inflow and outflow of water and sediment occurs at low rates. Discrete, high-energy flood events into or out of the lake have not occurred for a long period of time (e.g. 150 ka–1 Ma), allowing at least 6 m of high porosity, fine sediment to accumulate undisturbed at the lake bed. There are no indications that the lake was lower during this period, although we cannot say whether or not it has been higher than at present. In places on its SE side, SLE extends up to 500 m beyond the area initially determined solely from the radar data. None of our results suggest a better location for sampling the lake bed than that proposed by Woodward et al. (2010); however, even there, we cannot completely rule out the possible influence of mass movement on the sediment sequence.

Many different types of subglacial lakes exist, depending on glaciological and geological setting, as well as water through-flow and sedimentary characteristics. Our conclusions derive from just one setting so will not be universal; however, they are useful as a general overall guide, and are probably widely applicable at interior ice sheet locations.

7.4. Implications for subglacial lakes in general

In addition to those conclusions relating specifically to SLE, we make the following general conclusions which are probably applicable to all types of subglacial lake. When analysed for reflection strength, seismic data can yield more-comprehensive interpretations about a subglacial lake, than is possible from physiography alone, particularly in terms of the nature of the lacustrine characteristics and sedimentary environment. As well as giving bathymetry, seismic surveys can show whether or not soft sediments occur at the bed of a subglacial lake and also indicate the sediment type (coarse or fine grained). This can be critical information for understanding a lake's environment, for guiding the design of equipment to successfully retrieve sediment samples from the bed, and for their subsequent interpretation. Extending seismic surveys onto the adjacent grounded ice can show whether the surrounding area has a thawed, permeable bed, or a less-permeable bedrock (or frozen) one. This can indicate whether the lake itself is an open, or closed hydrological system, with implications for sediment and water input, as well as the potential for build-up of clathrates or dissolved gases (Siegert et al., 2012). Precise mapping of subglacial lake shorelines based on radar data alone, should be treated with caution. Using quantitative analysis of seismic data too can reduce the risk of misidentifying parts of the lake as grounded ice.

Mass movement events are to be expected, albeit infrequently, in subglacial lakes, particularly those with steep bed slopes. Even the central parts of a lake, previously expected to be isolated from slides (Siegert et al., 2012), could be affected by them. Sediment coring locations should avoid obvious mounds. Irrespective of core location, the possibility of disturbance by sediment mass movement should always be considered when interpreting recovered subglacial lake sediments.

Prior to any subglacial lake access at an ice sheet interior location, geophysical surveys, including seismic acquisition, are essential. They help to identify the optimum access location and contribute to the contextual basis on which to interpret any recovered samples. If necessary, geophysical data acquisition should be repeated immediately prior to (e.g. a few days) access drilling commencing. This will assess whether any changes, particularly mass movement events, have occurred since previous surveys.

Author contributions

Competing interests

The authors declare no competing interests.

Data availability

The datasets generated and analysed during this study are lodged with NERC's UK Polar Data Centre (www.bas.ac.uk/data/uk-pdc/) and are available from the corresponding author on request.

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References

Anderson, V.H., 1960. The Petrography of Some Rocks from Marie Byrd Land, Antarctica. Project 825, Report No. 2, Part VIII, The Ohio State University Research Foundation, Columbus, Ohio, pp. 1–27.

Bellwald, B., Hjelstuen, B.O., Sejrup, H.P., Halflidason, H., 2016. Postglacial mass movements and depositional environments in a high-latitude fjord system – Hardengerfjorden, Western Norway. Mar. Geol. 379, 157–175. https://doi.org/10.1016/j.margeo.2016.06.002.

Bentley, C.R., Kohnen, H., 1976. Seismic refraction measurements on internal friction in Antarctic ice. J. Geophys. Res. 81, 1519–1526.

Bentley, M.J., Christoffersen, P., Hodgson, D.A., Smith, A.M., Tulaczyk, S., Le Brocq, A.M., 2011. Subglacial Lake Sediments and Sedimentary Processes: Potential Archives of Ice Sheet Evolution, Past Environmental Change and the Presence of Life. Geophysical Monograph Series, vol. 192. AGU, Washington, DC, pp. 83–110.

Christoffersen, P., Tulaczyk, T., Wattrus, N.J., Peterson, J., Quintana-Krupinski, N., Clark, C.D., Sjunneskog, C., 2008. Large subglacial lake beneath the Laurentide Ice Sheet inferred from sedimentary sequences. Geology 36, 563–566. https://doi.org/10.1130/G24628A.1.

DeConto, R.M., Pollard, D., 2016. Contribution of Antarctica to past and future sea-level rise. Nature 531, 591–597. https://doi.org/10.1038/nature17145.

Domack, E., Duran, D., Leventer, A., Ishman, S., Doane, S., McCallum, S., Amblas, D., Ring, J., Gilbert, R., Prentice, M., 2005. Stability of the Larsen B ice shelf on the Antarctic Peninsula during the Holocene epoch. Nature 436, 681–685. https://doi.org/10.1038/nature03908.

Filina, I.Y., Lukin, V., Masolov, V., Blankenship, D.D., 2007. Unconsolidated Sediments from the Bottom of Lake Vostok from Seismic Data. USGS Open file report 2007-1047, Short research paper 031.

Gow, A.J., Epstein, S., Sheehy, W., 1979. On the origin of stratified debris in ice cores from the bottom of the Antarctic ice sheet. J. Glaciol. 23, 185–192.

Hamilton, E.L., 1971. Elastic properties of marine sediments. J. Geophys. Res. 76, 579–604.

Hamilton, E.L., 1979. Sound velocity gradients in marine sediments. J. Acoust. Soc. Am. 65, 909–922.

Haywood, A.M., Smellie, J.L., Ashworth, A.C., Cantrill, D.J., Florindo, F., Hambrey, M.J., Hill, D., Hillenbrand, C.-D., Hunter, S.J., Larter, R.D., Lear, C.H., Passchier, S., van der Wal, R., 2009. Middle Miocene to Pliocene history of Antarctic and the Southern Ocean. In: Florindo, F., Siegert, M.J. (Eds.), Antarctic Climate Evolution. In: Developments in Earth and Environmental Sciences, vol. 8. Elsevier B.V., pp. 401–463.

Hein, A.S., Woodward, J., Marrero, S.M., Dunning, S.A., Steig, E.J., Freeman, S.P.H.T., Stuart, F.M., Winter, K., Westoby, M., Sugden, D.E., 2016. Evidence for the stability of the West Antarctic Ice Sheet divide for 1.4 million years. Nat. Commun. 7, 10325. https://doi.org/10.1038/ncomms10325.

Hodgson, D.A., Bentley, M.J., Smith, J.A., Klepacki, J., Makinson, K., Smith, A.M., Shaw, K., Scherer, R., Powell, R., Tulaczyk, S., Rose, M., Pearce, D., Mowlem, M., Keen, P., Siegert, M.J., 2016. Technologies for retrieving sediment cores in Antarctic subglacial settings. Philos. Trans. R. Soc. A 374, 20150056. https://doi.org/10.1098/rsta.2015.0056.

Hodgson, D.A., Roberts, S.J., Bentley, M.J., Carmichael, E.L., Smith, J.A., Verleyen, E., Vyverman, W., Geissler, P., Leng, M.J., Sanderson, D.C.W., 2009. Exploring former subglacial Hodgson Lake. Paper II: palaeolimnology. Quat. Sci. Rev. 28, 2310–2325. https://doi.org/10.1016/j.quascirev.2009.04.014.

Hodgson, D.A., Verleyen, E., Squier, A.H., Sabbe, K., Keely, B.J., Saunders, K.M., Vyverman, W., 2006. Interglacial environments of coastal east Antarctica: comparison of MIS 1 (Holocene) and MIS 5e (Last Interglacial) lake-sediment records. Quat. Sci. Rev. 25, 179–197. https://doi.org/10.1016/j.quascirev.2005.03.004.

Hodson, T.O., Powell, R.D., Brachfeld, S.A., Tulaczyk, S., Scherer, R.P., the WISSARD Science Team, 2016. Physical processes in Subglacial Lake Whillans, West Antarctica: inferences from sediment cores. Earth Planet. Sci. Lett. 444, 56–63. https://doi.org/10.1016/j.epsl.2016.03.036.

Horgan, H.J., Anandakrishnan, S., Jacobel, R.W., Christianson, K., Alley, R.B., Heeszel, D.S., Picotti, S., Walter, J.I., 2012. Subglacial Lake Whillans – seismic observations of a shallow active reservoir beneath a West Antarctic ice stream. Earth Planet. Sci. Lett. 331–332, 201–209. https://doi.org/10.1016/j.epsl.2012.02.023.

Jankowski, E.J., Drewry, D.J., 1981. The structure of West Antarctica from geophysical studies. Nature 291, 17–21.

Kanao, M., 2014. Seismicity in the Antarctic continent and surrounding ocean. Open J. Earthq. Res. 3, 5–14. https://doi.org/10.4236/ojer.2014.31002.

Kremer, K., Hilbe, M., Simpson, G., Decrouy, L., Wild, W., Girardclos, S., 2015. Reconstructing 4000 years of mass movement and tsunami history in a deep peri-Alpine lake (Lake Geneva, France–Switzerland). Sedimentology 62, 1305–1327. https://doi.org/10.1111/sed.12190.

Kuhn, G., Hillenbrand, C-D., Kasten, S., Smith, J.A., Nitsche, F.O., Frederichs, T., Wiers, S., Ehrmann, W., Klages, J.P., Mogollón, 2017. Evidence for a palaeo-subglacial lake on the Antarctic continental shelf. Nat. Commun. 8, 15591. https://doi.org/10.1038/ncomms15591.

LeBlanc, L.R., Mayer, L., Rufino, M., Schock, S.G., King, J., 1992. Marine sediment classification using chirp sonar. J. Acoust. Soc. Am. 9, 107–115. https://doi.org/10.1121/1.402758.

Lukin, V.V., Vasiliev, N.I., 2014. Technological aspects of the final phase of drilling borehole 5G and unsealing Vostok Subglacial Lake, East Antarctica. Ann. Glaciol. 55, 83–89. https://doi.org/10.3189/2014AoG65A002.

Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., Pekar, S.F., 2005. The Phanerozoic record of global sea-level change. Science 310, 1293–1298. https://doi.org/10.1126/science.1116412.

Pattyn, F., 2010. Antarctic subglacial conditions inferred from a hybrid ice sheet/ice stream model. Earth Planet. Sci. Lett. 295, 451–461. https://doi.org/10.1016/j.epsl.2010.04.025.

Peters, L.E., Anandakrishnan, S., Holland, C.W., Horgan, H.J., Blankenship, D.D., Voigt, D.E., 2008. Seismic detection of a subglacial lake near the South Pole, Antarctica. Geophys. Res. Lett. 35, L23501. https://doi.org/10.1029/2008GL035704.

Pollard, D., DeConto, R.M., 2009. Modelling West Antarctic ice sheet growth and collapse through the past five million years. Nature 458. https://doi.org/10.1038/nature07809.

Richardson, M.D., Briggs, K.B., 1993. On the use of acoustic impedance values to determine sediment properties. Proc. Inst. Acoust. 15, 15–24.

Rivera, A., Sharp, M., McDowell, P., Zamora, R., Neira, G., Wendt, J., Wendt, A., Araya, L., Ulloa, D., Rodríguez, M., Acuña, C., Arévalo, M., Casassa, G., Ordenes, F., Siegert, M., Bown, F., 2006. Expedición al lago subglacial Ellsworth. Bol. Antárt. Chil. 25, 7–10.

Ross, N., Jordan, T.A., Bingham, R.G., Corr, H.F.J., Ferraccioli, F., Le Brocq, A., Rippin, D.M., Wright, A.P., Siegert, M.J., 2014. The Ellsworth Subglacial Highlands: inception and retreat of the West Antarctic Ice Sheet. Geol. Soc. Am. Bull. 126, 3–15. https://doi.org/10.1130/B30794.1.

Ross, N., Siegert, M.J., Rivera, A., Bentley, M.J., Blake, D., Capper, L., Clarke, R., Cockell, C.S., Corr, H.F.J., Harris, W., Hill, C., Hindmarsh, R.C.A., Hodgson, D.A., King, E.C., Lamb, H., Maher, B., Makinson, K., Mowlem, M., Parnell, J., Pearce, D.A., Priscu, J., Smith, A.M., Tait, A., Tranter, M., Wadham, J.L., Whalley, W.B., Woodward, J., 2011b. Ellsworth Subglacial Lake, West Antarctica: A Review of Its History and Recent Field Campaigns. Geophysical Monograph Series, vol. 192. AGU, Washington, DC, pp. 221–233.

Ross, N., Siegert, M.J., Woodward, J., Smith, A.M., Corr, H.F.J., Bentley, M.J., Hindmarsh, R.C.A., King, E.C., Rivera, A., 2011a. Holocene stability of the Amundsen–Weddell ice divide, West Antarctica. Geology 39, 935–938. https://doi.org/10.1130/G31920.1.

Schnellmann, M., Anselmetti, F.S., Giardini, D., McKenzie, J.A., 2006. 15,000 years of mass-movement history in Lake Lucerne: implications for seismic and tsunami hazards. Eclogae Geol. Helv. 99, 409–428. https://doi.org/10.1007/S00015-006-1196-7.

Siegert, M.J., Clarke, R.J., Mowlem, M., Ross, N., Hill, C.S., Tait, A., Hodgson, D., Parnell, J., Tranter, M., Pearce, D., Bentley, M.J., Cockell, C., Tsaloglou, M.-N., Smith, A., Woodward, J., Brito, M.P., Waugh, E., 2012. Clean access, measurement and sampling of Ellsworth Subglacial Lake: a method for exploring deep Antarctic subglacial lake environments. Rev. Geophys. 50, RG1003. https://doi.org/10.1029/2011RG000361.

Siegert, M.J., Makinson, K., Blake, D., Mowlem, M., Ross, N., 2014. An assessment of deep hot-water drilling as a means to undertake direct measurement and sampling of Antarctic subglacial lakes: experience and lessons learned from the Lake Ellsworth field season 2012/13. Ann. Glaciol. 55, 59–73. https://doi.org/10.3189/2014AoG65A008.

Siegert, M.J., Ross, N., Le Brocq, A., 2016. Recent advances in understanding Antarctic subglacial lakes and hydrology. Philos. Trans. R. Soc. A 374, 20140306. https://doi.org/10.1098/rsta.2014.0306.

Smith, A.M., 1997. Basal conditions on Rutford Ice Stream West Antarctica, from seismic observations. J. Geophys. Res. 102, 543–552.

Smith, A.M., Jordan, T.A., Ferraccioli, F., Bingham, R.G., 2013. Influence of subglacial conditions on ice stream dynamics: seismic and potential field data from Pine Island Glacier, West Antarctica. J. Geophys. Res. 118, 1–12. https://doi.org/10.1029/2012JB009582.

Smith, A.M., Murray, T., Davison, B.M., Clough, A.F., Woodward, J., Jiskoot, H., 2002. Late surge glacial conditions on Bakaninbreen, Svalbard, and implications for surge termination. J. Geophys. Res. 107. https://doi.org/10.1029/2001JB000475.

Smith, B.E., Fricker, H.A., Joughin, I.R., Tulaczyk, T., 2009. An inventory of active subglacial lakes in Antarctica detected by ICESat (2003–2008). J. Glaciol. 55, 573–595.

Stearns, L.A., Smith, B.E., Hamilton, G.S., 2008. Increased flow speed on a large East Antarctic outlet glacier caused by subglacial floods. Nat. Geosci. 1, 827–831. https://doi.org/10.1038/ngeo356.

Syvitski, J.P.M., Hein, F.J., 1991. Sedimentology of an Arctic Basin: Itirbilung Fjord, Baffin Island, Northwest Territories. Geological Survey of Canada Professional Paper 91-11.

Syvitski, J.P.M., Shaw, J., 1995. Sedimentology and geomorphology of fjords. In: Perillo, G.M.E. (Ed.), Geomorphology and Sedimentology of Estuaries. In: Developments in Sedimentology, vol. 53. Elsevier Science B. V., pp. 113–178.

Thoma, M., Grosfeld, K., Mayer, C., Smith, A.M., Woodward, J., Ross, N., 2011. The "tipping" temperature within Subglacial Lake Ellsworth, West Antarctica and its implications for lake access. Cryosphere 5, 561–567. https://doi.org/10.5194/tc-5-561-2011.

Tulaczyk, S., Mikucki, J.A., Siegfried, M.R., Priscu, J.C., Barcheck, C.G., Beem, L.H., Behar, A., Burnett, J., Christner, B.C., Fisher, A.T., Fricker, H.A., Mankoff, K.D., Powell, R.D., Rack, F., Sampson, D., Scherer, R.P., Schwartz, S.Y., the WISSARD Science Team, 2014. WISSARD at Subglacial Lake Whillans, West Antarctica: scientific operations and initial observations. Ann. Glaciol. 55, 51–58. https://doi.org/10.3189/2014AoG65A009.

Vardy, M.E., 2015. Deriving shallow-water sediment properties using post-stack acoustic impedance inversion. Near Surf. Geophys. 13, 143–154. https://doi.org/10.3997/1873-0604.2014045.

Vaughan, D.G., Rivera, A., Woodward, J., Corr, H.F.J., Wendt, J., Zamora, R., 2007. Topographic and hydrological controls on Subglacial Lake Ellsworth, West Antarctica. Geophys. Res. Lett. 34, L18501. https://doi.org/10.1029/2007GL030769.

Wingham, D.J., Siegert, M.J., Shepherd, A., Muir, A.S., 2006. Rapid discharge connects Antarctic subglacial lakes. Nature 440, 1033–1036. https://doi.org/10.1038/nature04660.

Woodward, J., Smith, A.M., Ross, N., Thoma, M., Corr, H.F.J., King, E.C., King, M.A., Grosfeld, K., Tranter, M., Siegert, M.J., 2010. Location for direct access to subglacial Lake Ellsworth: an assessment of geophysical data and modelling. Geophys. Res. Lett. 37, L11501. https://doi.org/10.1029/2010GL042884.