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Antarctic subglacial lakes

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Abstract

Antarctic subglacial lakes were first identified by Robin et al. (1970) after airborne radio-echo sounding (RES) investigations of the ice-sheet interior. Recently, satellite altimetry was used to measure anomalous near-flat regions on the ice-sheet surface that represent a manifestation of the subglacial lake beneath. Using RES and satellite altimetry, the location and extent of Antarctic subglacial lakes can be identified. The largest subglacial lake exists beneath Vostok Station, and is 14,000 km^2 in area. The combined area of additional subglacial lakes beneath Dome C is 15,000 km^2 and at least 15,000 km^2 of lake surface lies beneath the remainder of the ice sheet. The water depth of subglacial lakes can be estimated through seismic investigations (although data exist only for Lake Vostok) and consideration of the bedrock slopes that border subglacial lakes. The depths of many subglacial lakes are of the order of 10's–100's of metres. The total volume of water held beneath the ice sheet is estimated between 4000 and 12,000 km^3. To date, there are six known examples of radio-echo reflections from the lake floors (at a depth of no more than 20 m). Since e/m attenuation through water is related to the salinity, these data indicate that subglacial water is very pure and fresh. Some near-flat surface regions that usually occur over lakes have been observed where no lakes exist. Such features are may be caused by water-saturated basal sediments rather than subglacial lakes. Finally, the spatial variation in geothermal heat flux around the central regions of Antarctica can be established estimated by employing a simple thermal model of the ice sheet under an assumption that the basal ice temperature above subglacial lakes is equal to the pressure melting value. Calculations indicate that the geothermal heat flux varies spatially over the Antarctic Plate between 37 and 65 mW m^{-2}.

Keywords: Antarctica; subglacial; lakes; glaciology; ice sheet

1. Introduction and background

1.1. The discovery of Antarctic subglacial lakes

Subglacial lakes are discrete bodies of water that lie at the base of an ice sheet between ice and substrate. Antarctic subglacial lakes were first discovered over 20 years ago. The first, rather unwitting, observation of a subglacial lake was by Robinson (1960) who, as part of an experiment to determine ice-surface landmarks to aid flight orientation, identified ''oval depressions with gentle shores'' on the ice surface. Although these features were referred to as 'lakes' by pilots who observed them, such as Robinson, there was no connection made

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between ice-surface morphology and basal ice-sheet conditions. However, during the 1990s, as will be discussed later, a relationship between surface ice-sheet morphology and subglacial lakes was established.

The first direct identification of a subglacial lake came as a consequence of the Antarctic airborne radio-echo sounding (RES) program between 1968 and 1979 (referred to later). Radio-echo sounding is capable of acquiring information about the ice-sheet base and so is instrumental in the identification of basal water bodies. During one of the first RES flights across the Antarctic continent a 10-km-long RES signal, representing an unusually flat subglacial surface, was received beneath the Russian base at Sovetskaya Station in central East Antarctica (Robin et al., 1970). This signal was attributed to a ‘‘thick layer of water beneath the ice’’ (Robin et al., 1970). Subsequent to this discovery, RES was used to identify a number of other Antarctic subglacial lakes (Oswald and Robin, 1973; McIntyre, 1983). There have been few subsequent investigations until relatively recently. In the last few years a re-examination of a large RES database has revealed new information about their size, occurrence and glacial and geological significance of Antarctic subglacial lakes.

The aim of this paper is to summarisze what is currently understood about the physical characteristics of Antarctic subglacial lakes. The influence that subglacial lakes exert on ice-sheet dynamics is discussed with reference to satellite altimetry of the ice surface. The geological significance of subglacial lakes in terms of stores for continuous sediment build-up, and sources for water outburst events are then discussed.

1.2. A brief introduction to the Antarctic Ice Sheet

Antarctica is comprised of two main grounded ice sheets; the West Antarctic and the East Antarctic ice sheets. The present volume of ice in Antarctica is around 30,000 km^3 (equivalent to 60 m of global sea-level) of which 83% is housed within the East Antarctic Ice Sheet. The maximum surface elevation is just over 4000 m in East Antarctica. Ice thickness varies between 2800 and 4500 m in the central regions of Antarctica. The ice thickness and bedrock elevation (at the base of the ice sheet) can vary spatially by 100s of metres over a few km. The basal topography is arranged as a series of troughs, basins and highlands. Subglacial topography is fundamental to the location of subglacial lakes since lakes are generally located within topographic hollows. The East and West Antarctic ice sheets are separated by the Trans-Antarctic mountains. If the ice were to be removed from East Antarctica, the bedrock surface would be above sea-level. However, if the same were to happen over West Antarctica, even accounting for isostatic rebound, the bedrock would remain below the modern sea level. Thus, the West Antarctic Ice Sheet is often referred to as a marine-based ice sheet. In addition to the two main ice sheets, the other significant region of grounded ice exists over the Antarctic Peninsular where a series of ice caps and glaciers are located.

Grounded ice flows by (1) the deformation of ice, and (2) basal sliding over bedrock and/or (3) deformation of water saturated basal substrate. Ice is drained from the ice domes of the ice-sheet interior by fast flowing rivers of ice known as ‘ice streams’, where the dominant method of flow is by sliding and/or basal sediment deformation. These feed grounded ice into numerous floating ‘ice shelves’ that surround the continent. The largest two ice shelves are the Ross and the Filchner-Ronne ice shelves which have an area of 526,000 and 473,000 km^2, respectively (Drewry, 1983). Icebergs, usually of tabular form, are calved out from the marine margins of ice shelves, where ice thickness is usually around 250–300 m. This process represents the most important mechanism by which ice is lost from the ice-sheet system; accounting for 85–75% of ice loss (Jacobs et al., 1992; Paterson, 1994). A full review of the Antarctic Ice Sheet, and methods by which ice sheets flow can be found in *Quaternary Science Reviews* Vol. 9 (1990) and Paterson (1994), respectively.

2. Identification, occurrence and dimensions of subglacial lakes

2.1. The identification of subglacial lakes

An extensive RES database is held in archive at the Scott Polar Research Institute (SPRI), University of Cambridge. These data were collected between 1968 and 1979 as part of the collaborative airborne RES survey of the ice sheet between the SPRI and the Technical University of Denmark (TUD) and the National Science Foundation (NSF) of the USA. The SPRI-TUD-NSF survey covers, through 400,000 km of flight track, around 50% of East Antarctica, as well as the Ross Ice Shelf and the southern most 30% of West Antarctica covering a total area of ~ 7 million km^2. Since the acquisition of SPRI-TUD-NSF RES data, there have been numerous smaller RES surveys over the some of the remaining regions of Antarctica. However, a large part of Antarctica has yet to be surveyed by this method. Unless stated otherwise, in this paper information from subglacial lakes has been gained from the SPRI RES database. Airborne RES works by the emission of a radio wave which is reflected off boundaries of dielectric contrast and received again very close to the location of transmission. The one-way travel time of the radio-wave can be multiplied by the velocity of e/m radiation through the propagating material to yield the total distance travelled. The radio-wave velocity through ice is 168 m ms^{-1} (e.g., Glen and Paren, 1975). This calculation yields information about ice thickness with an accuracy of about 1.5%. By recording depth sequentially every few seconds as the aircraft traverses across the ice sheet, a pseudo cross-section of the ice sheet can be obtained (Fig. 1). For the surveys performed prior to availability of digital technology, the received radio-wave was recorded onto 35 mm film from which hardcopies of the RES data were produced. Through photographic processing, the horizontal axis (i.e., flight distance) could be compressed so that vertically exaggerated cross-sections that comprise several 10's to 100's of flight track can be viewed. It should be noted that modern RES equipment record digital information to computer disks, on which processing can be performed using image analysis software.

Airborne RES at 60 MHz has been used successfully to penetrate to the base of ice over 4 km thick in Antarctica (e.g., Robin et al., 1977; Drewry, 1983). This is possible because ice is relatively transparent to e/m radiation at this frequency (Johari and Charette, 1975), especially when ice is several tens of degrees below freezing, as is the case for most of the Antarctic Ice Sheet.

The strength of reflection from the bed depends predominantly upon the difference between the dielectric properties of the ice (real part of the dielectric constant $\varepsilon = 3.2$), and the dielectric properties of the sub-ice material (i.e., a boundary of dielectric contrast). As the dielectric constant of water ($\varepsilon = 81$) is very different from typical bedrock ($\varepsilon = 4$ to 9), a much stronger reflection is obtained from an ice-water interface compared with an ice-rock interface. This difference is further increased by the relatively rough character of an ice-bedrock interface, which scatters energy and further reduces echo strength.

Sub-glacial lakes are identified on 60 MHz RES records by the presence of the three distinct characteristics (Fig. 1). The first is especially strong reflections from the ice-sheet base, which appear bright on 35 mm film records and are typically 10–20 dB stronger than adjacent ice-bedrock reflections. The second characteristic is that echoes of constant strength occur along the track, indicative of an interface which is very smooth on the scale of the RES wavelength. The third is a very flat and virtually horizontal character (i.e., mirror-like), with maximum slopes typically less than 1%.

Such subglacial lake radio-wave reflections are highly distinctive and pose relatively few problems concerning identification from the surrounding ice-bedrock interface (Fig. 1). However, it should be noted that accumulations of water-saturated basal sediments may also yield relatively strong radio-wave returns, in some cases similar to those observed from an ice-water interface. Conceivably, such returns may be observed on the RES record as flat horizontal reflections. This possibility is referred to later in a section on the geological significance of subglacial lakes.

2.2. The spatial distribution of Antarctic subglacial lakes

RES records of the Antarctic ice sheet are linked to aircraft navigational information. Because of this, the location of subglacial lake RES data can be determined. In addition, the length of the water body and the thickness of overlying ice has also been measured from the RES data. Lengths of the sub-ice lake records were calculated by assuming a constant aircraft speed of 300 km h^{-1}, and should be consid-

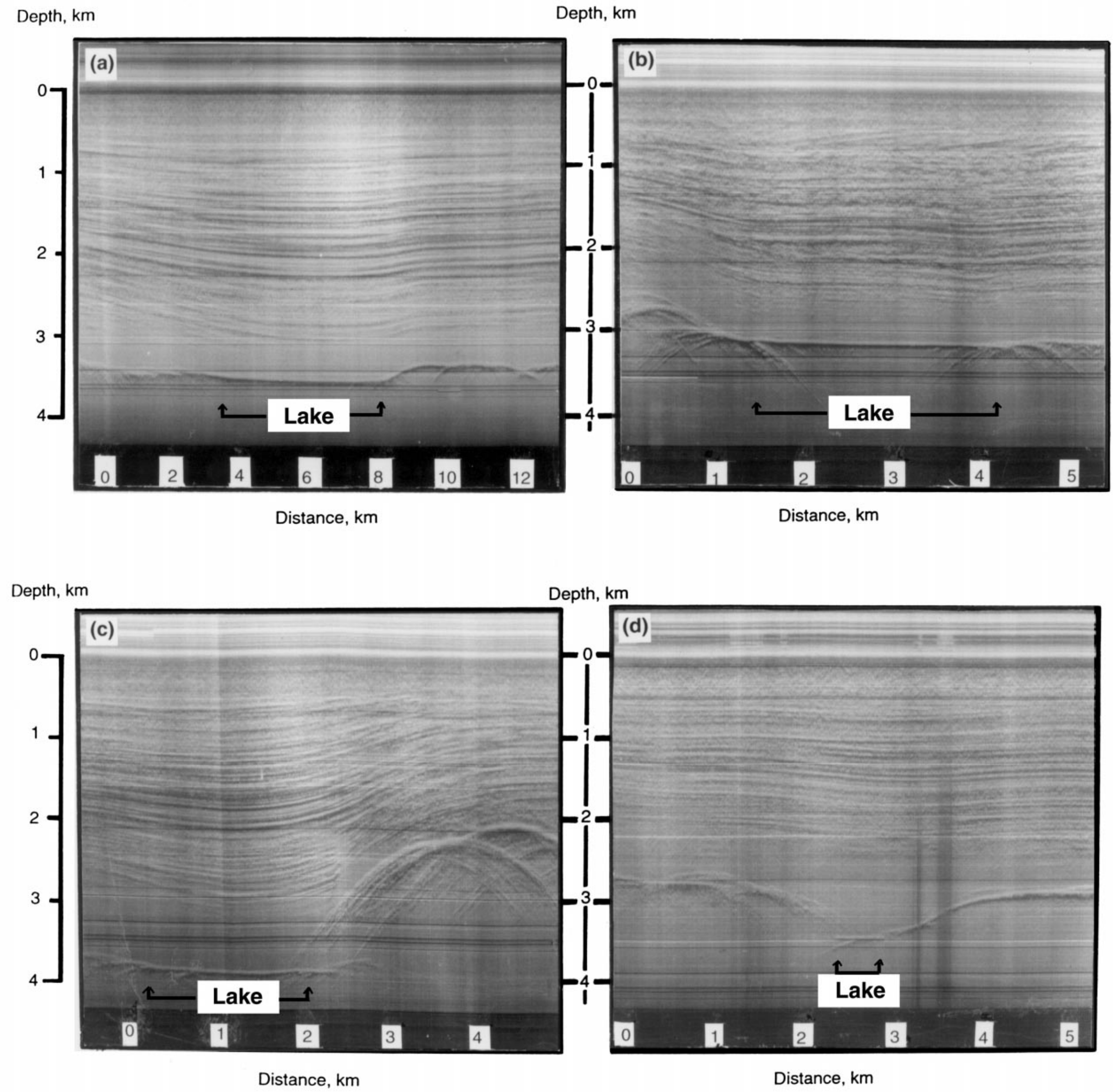


Fig. 1. Typical 60 MHz radio-echo sounding records of four small Antarctic sub-glacial lakes. Note the strong, flat reflections which are used to identify lakes, as compared with the weaker, undulating signals reflected off the surrounding subglacial bedrock (adapted from Dowdeswell and Siegert, 1999).

ered as representing minimum values for water-body dimensions (Siegert et al., 1996).

Siegert et al. (1996) identified 77 sub-glacial lake-type RES reflectors. The majority of the observed lakes are situated in relatively close proximity to ice divides, where both the surface slope and ice velocity are small (Fig. 2). Two clusters of lakes, accounting for 70% of the total lake inventory, are in regions of Dome C and Ridge B in East Antarctica. In addition, the ice divide stretching from West to

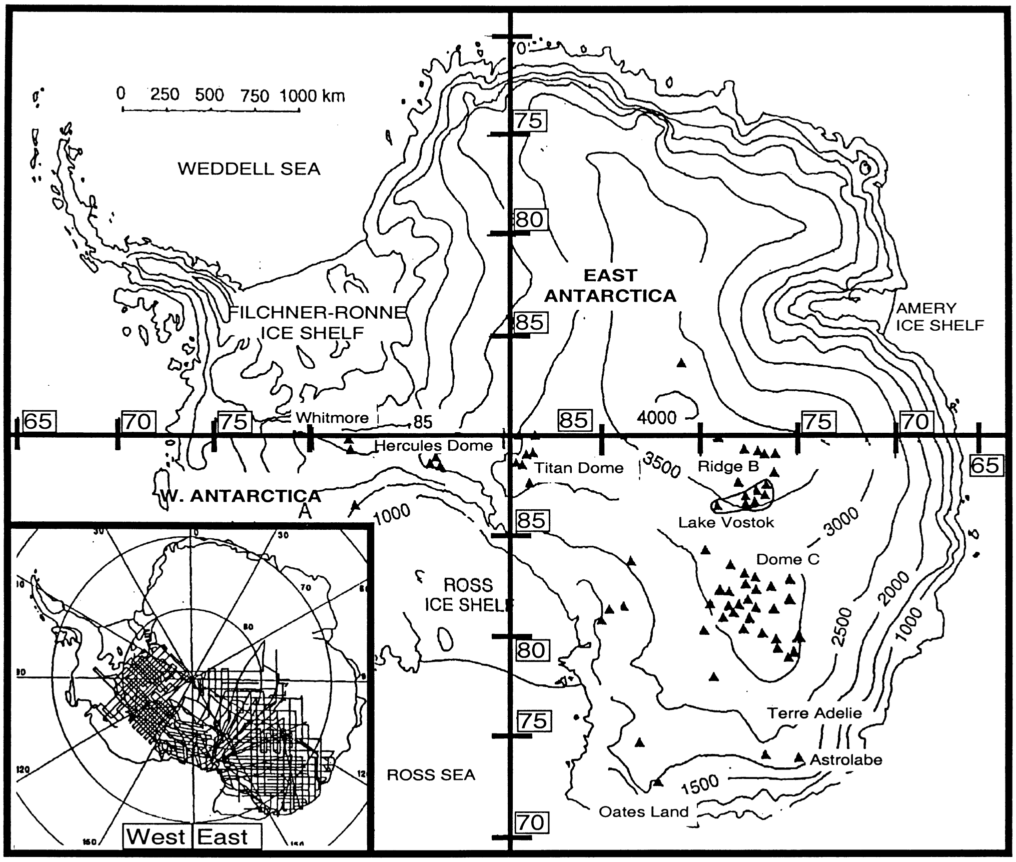


Fig. 2. The locations of lakes beneath the Antarctic Ice Sheet (marked by triangles), identified from 60 MHz airborne radio-echo sounding records. Ice-sheet surface elevation is contoured at 500 m intervals (adapted from Siegert et al., 1996).

East Antarctica (which runs close to the South Pole), has several sub-ice lakes along its length in the areas of Hercules Dome and Titan Dome (Fig. 2). At least 16 subglacial lakes are located away from the ice divide, close to the onset of enhanced ice-sheet flow.

It is likely that the total number of lakes identified by Siegert et al. (1996) is an underestimate. First, flight-line spacing of 50-100 km over much of the SPRI-TUD-NSF survey area means that some additional lakes may be present between existing flight-lines. For example, in a recent RES survey of the central region of Dome C (Fig. 2), performed in order to determine a location for the new EPICA ice core site, two small subglacial lakes were discovered between one block of flight tracks from the SPRI-TUD-NSF survey (Tabacco et al., 1998). Second, where ice above a subglacial lake has been subject to basal sliding, the ice-sheet base may include an imprint of the rough upstream bedrock morphology. In this situation, RES may not record the ice-water interface as a ‘mirror like’ reflector and, thus, not detect the lake (Reynolds and Whillans, 1979). Third, several regions of Antarctica have yet to be surveyed by the RES method. Therefore, it is highly likely that a number of subglacial lakes exist undetected beneath the Antarctic Ice Sheet.

Using ice-surface topography and flow lines reconstructed for the modern Antarctic Ice Sheet from surface altimetric data (Drewry, 1983), the distance of each lake from the ice divide can be calculated. Several sub-glacial lakes occur on ice divides, and almost 40% are located within 100 km of the ice crest in drainage basins which are often about 1000 km from divide to seaward margin (Fig. 3). The mean thickness of ice above the lakes is about 3000 m, with a standard deviation of 460 m. Less than 5% of sub-glacial lakes occur beneath ice of less than 2000 m in thickness. This is consistent with their tendency to be located close to the thick ice associated with ice divides.

2.3. The dimensions of subglacial lakes from RES data

At each lake site identified from RES records, the horizontal length of the strong, smooth basal reflector has also been measured. The longest lake is 230 km and is located close to Vostok Station in East Antarctica (Kapitsa et al., 1996). Seventy-five per-

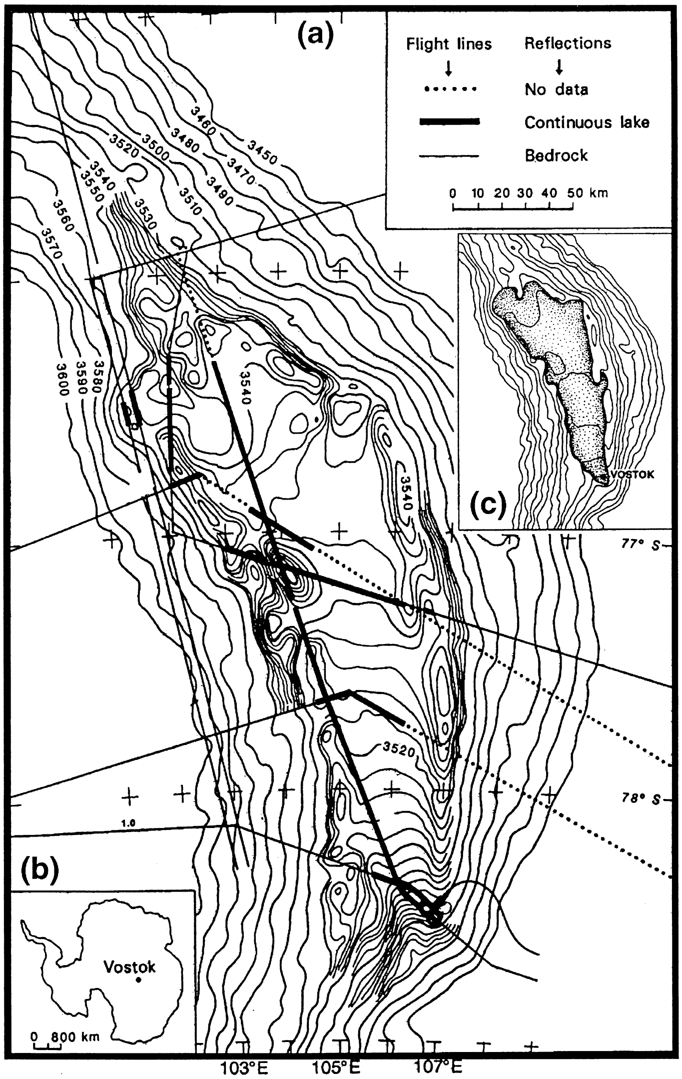


Fig. 3. (a) ERS-1 satellite radar altimeter map of the Lake Vostok area. Contours are every 10 m. The positions of all RES flight lines over the lake are provided, and regions of 'bedrock reflection' and 'lake reflection' noted. The diagram indicates that the lake lies beneath Vostok Station, and extends northwards for 230 km. (b) Location of Vostok Station. (c) Location and margin of Lake Vostok, inferred from RES data and ERS-1 altimetry (adapted from Kapitsa et al., 1996). (d) RES flightlines around the Vostok Station magnified from (a). Adapted from Siegert and Ridley (1998b).

cent of lakes have radio-echo lengths of less than 10 km, and only 5% are greater than 30 km. The mean observed lake length is 10.8 km, excluding Lake Vostok. However, it should be stressed that these observations represent *minimum* lake lengths. It is unlikely that radio-echo flight-lines passed over the longest axis of all but a small number of lakes, and in many cases a single line rather than a network of lines identified a given lake.

2.4. Satellite altimetry as an indicator of subglacial lake size

Satellite radar altimetery detailing the surface of the Antarctic Ice Sheet can be used to investigate the dimensions of sub-glacial lakes. The frictionless contact between ice and water at the ice-sheet base above a subglacial lake manifests itself as an anomalous near-flat region on the ice-sheet surface (as witnessed by Robinson, 1960). Comparison of the extensive area of very nearly flat ice around Vostok Station with RES data demonstrates that the extent of larger lakes can be identified from satellite altimetry (Ridley et al., 1993; Kapitsa et al., 1996). No other lakes of similar magnitude to Lake Vostok have been reported from altimetric data of the Antarctic Ice Sheet below their ERS-1 latitudinal limit of 82°S (J.K. Ridley, pers. comm.). However, the surface areas for a further 28 sub-glacial lakes have been determined from satellite altimetry of the Dome C and Terre Adélie regions (Cudlip and McIntyre, 1987; Siegert and Ridley, 1998a). Small subglacial lakes of less than about 4 km across cannot be identified using radar altimetry because there will be no observable flat topography at the ice surface some 3 km above (Siegert and Ridley, 1998a). No satellite altimeter data exist for lakes around South Pole or Hercules Dome, due to the absence of satellite coverage in these latitudes. The surface areas of such lakes must be estimated from RES data alone.

2.5. Lake Vostok

2.5.1. Extent of the lake

The extent of the Vostok subglacial lake can be estimated by the comparison of two geophysical datasets (1) airborne RES and (2) satellite altimetry (from, for instance, the ERS-1 satellite) of the flat region on the ice-sheet surface above the subglacial lake.

To date there has been no systematic attempt to map Lake Vostok by RES methods. Consequently, data from only a few flights lines are available for the region. Eight RES flight lines which traversed Lake Vostok during the 1970s provide RES information about the thickness of the ice sheet over, and the extent of, the lake. Fig. 3 illustrates the flight paths, and the nature of the ice-sheet base determined from the RES data.

The ERS-1 was the first satellite equipped with a radar altimeter to cover the Antarctic plateau up to 82°S. Through the application of corrections to the height estimate and the best available satellite orbit data, an individual satellite-derived height measurement over the plateau has a precision of 25 cm and an accuracy of about 1 m (Siegert and Ridley, 1998b). The ERS-1 satellite produced a detailed surface elevation map of the ice sheet above Lake Vostok (Fig. 3) (Kapitsa et al., 1996; Siegert and Ridley, 1998b). There is a very strong correlation between the RES-determined location of Lake Vostok (Robin et al., 1977; Ridley et al., 1993; Kapitsa et al., 1996) and the position of an anomalous flat-surfaced region of the ice sheet (Fig. 3). From the margin of the flat surface region, since it matches so well with the RES-observed edge of the lake, the lake extent has been mapped (Fig. 3c). Lake Vostok is thus 230 km long, and up to 50 km wide, with an estimate surface area of 14,000 km^2 making it one of the world's largest lakes (Lake Vostok is 20 times the size of Lake Geneva and one half that of Lake Baykal). Satellite radar altimetry shows that the ice sheet above Lake Vostok has an approximately north-south surface slope of 0.004° (0.25 m km^{-1}) (Fig. 3a). The ice-sheet surface thus drops by about 40 m across the 230 km length of the lake.

Through analysis of the RES and ERS-1 data, representations of the ice thickness and basal topography have been established (Fig. 4). Ice thickness is more than 4 km over the northern region of the lake, and thins steadily along the length of the lake such that, beneath Vostok Station, it is around 3.5 km. The ice thickness surrounding the lake is generally less than 4000 m, where the bedrock elevation is

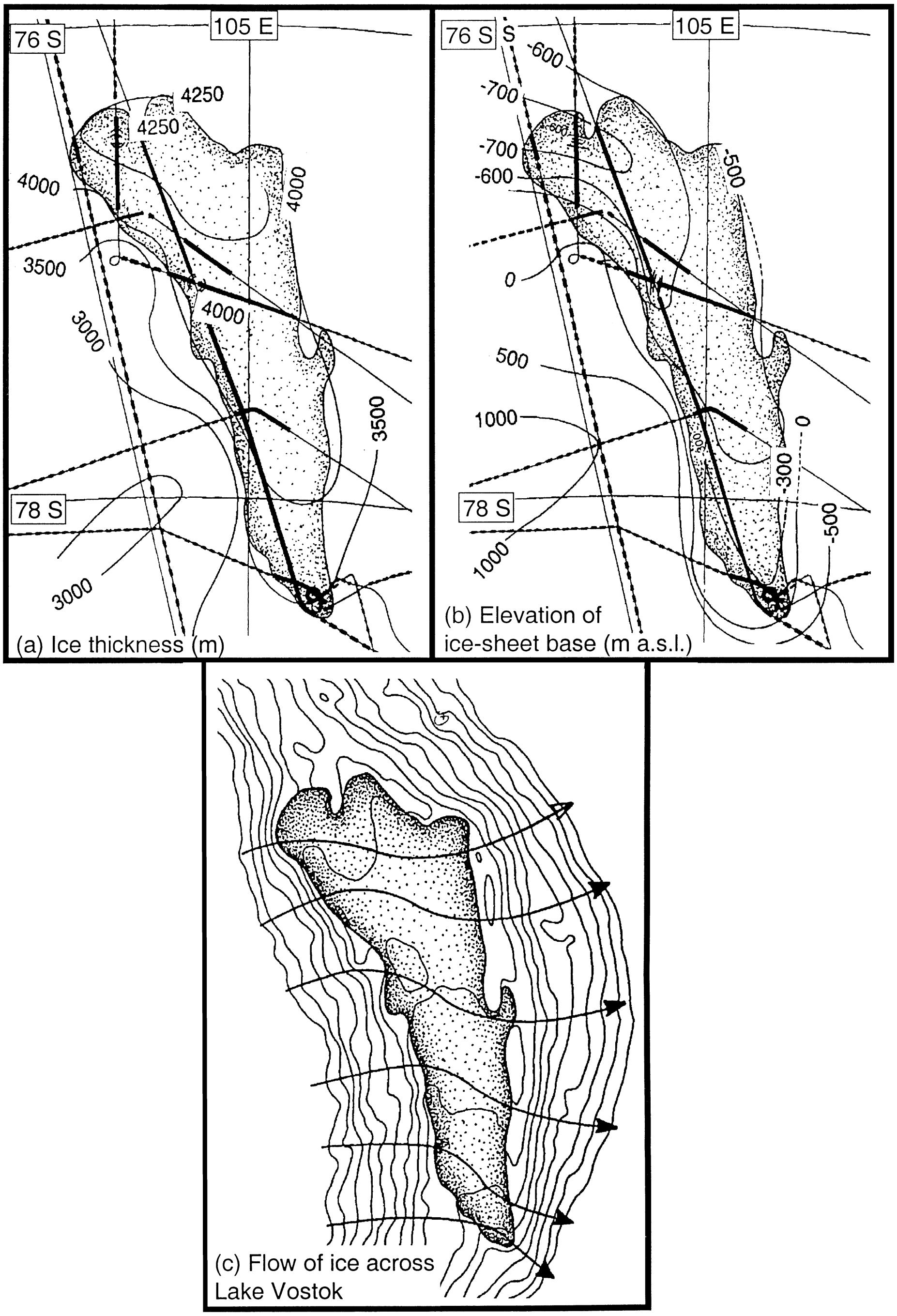


Fig. 4. (a) Ice thickness map of the Lake Vostok region. (b) Map of the elevation of the ice-sheet base, around the Lake Vostok region. (c) The likely flow of ice over Lake Vostok. The direction of ice flow is controlled by both the surrounding easterly, and local southerly, ice surface slopes. Adapted from Siegert and Ridley (1998b).

high relative to the lake basin. The height of the lake surface at its northern end is around −700 m above sea level (Fig. 4). This elevation increases toward the south of the lake where, beneath Vostok Station, the surface elevation of the lake is at about −300 m.

2.5.2. Water depth of Lake Vostok

Information about the depth of Lake Vostok comes from three sources: (1) seismic investigations; (2) RES-derived information about the slopes of bedrock which border the lake; and (3) reflection of 60 MHz RES waves off the lake floor.

Seismic surveys are a means to obtain point measurements of ice thickness and, because the passage of seismic p-waves are unimpeded by water, the depth of water within subglacial lakes. The only available seismic information for Lake Vostok is from experiments performed 1 km Northwest of Vostok Station (e.g., Kapitsa et al., 1996). Fig. 5 shows the relevant section of a seismic record using 10 seismometers space vertically along a borehole between 2.5 and 49 m depth instead of the usual horizontal spread. From a shot depth of 33 m, a primary echo is seen at 1.89 s and is followed by associated echoes to 2.00 s as p-waves travel up and down the seismometer line, having been reflected also at the ice-sheet surface. A similar pattern of echoes between 2.62 and 2.73 s show deeper p-wave reflections arriving vertically at the borehole from the lake bed. Taking the sound velocity in water as 1450 m s^{-1}, the time interval of $\sim$0.7 s between the first arrival from the ice–water interface and the lake bed, indicates a lake depth of $\sim$500 m.

RES information on the ice-sheet basal topography shows that Lake Vostok is located within a subglacial trough (Fig. 4b). Although RES can provide no explicit information concerning the depth of water bodies deeper than a few metres (Drewry, 1983; Gorman and Siegert, 1999), RES data from the side slopes bordering Lake Vostok can be used to estimate the depth of water. The steepest basal slope observed is along flight 123, where the bedrock elevation falls by more than 1000 m over a horizontal distance of approximately 10 km, a gradient of 0.1. If this gradient continues beneath the lake then the mean lake depth would be of the order of several hundred metres (Siegert and Ridley, 1998b).

RES reflections from the floor of six sub-glacial lakes, including the northern region of Lake Vostok, have been identified (Gorman and Siegert, 1999). Information regarding the depth of these lakes can be gained by measuring the travel time of the secondary reflections, since the velocity of e/m waves in water (33 m μs^{-1}) is relatively unaffected by salinity. Results from this exercise indicates that the water depths of these sub-glacial lakes vary between 10 and 20 m. Because of the absorption of radio-wave energy in water, subglacial lakes deeper than about 20 m will not be detectable by this method.

Combination of these three datasets suggests that the average depth of the Lake Vostok may will be several hundred metres in its southern end, but only a few tens of meters in the northern region. A comparable topographic feature to the trough in which Lake Vostok is located is the Astrolabe subglacial basin within Terre Adélie, which has been

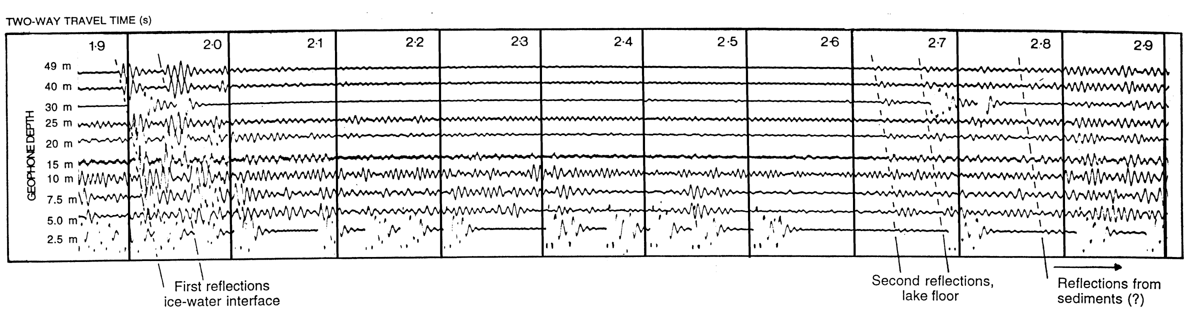


Fig. 5. Results from a seismic experiment located a few km from Vostok Station. Geophones are located down a borehole to a depth of 49 m. P-wave returns indicate two significant reflections, one from the ice-sheet base at 1.95 s and another 0.7 s later from the water–substrate interface at the floor of Lake Vostok (Taken from Kapitsa et al., 1996).

likened to the Great Rift Valley in East Africa (Cudlip and McIntyre, 1987; Siegert and Glasser, 1997).

2.5.3. Volume of water in Lake Vostok

With an area around 14,000 km^2, an envelope for the water volume held within Lake Vostok, based on information relating to water depth, can be estimated. For example, if the 510 m depth represented an average depth, then the volume of Lake Vostok would be 7140 km^3. However, this is probably a maximum estimate. A more realistic estimate, based on consideration of bedrock slopes at the lake margins, is an average water depth of 200 m, giving a water volume of 2800 km^3. A minimum estimate would be an average water depth of 100 m, yielding a water volume of 1400 km^3. Thus, the volume of water within lake Vostok lies within a likely envelope between 7000 and 1500 km^3. Kapitsa et al. (1996) estimated that the water volume of Lake Vostok is most likely to be 1800 km^3.

3. Glaciological significance of Antarctic subglacial lakes

3.1. Ice sheet topography above subglacial lakes

As is the case for Lake Vostok, many other subglacial lakes manifest themselves as relatively flat regions on the ice-sheet surface (compared with the general slope of surrounding ice). This is because shear stress at the ice-sheet basal interface is reduced to zero when over the subglacial lake, and glacial dynamics changed from the base-parallel shearing of grounded ice to longitudinal extension when ice is floating (as in ice shelves).

As a consequence, the association between ice surface morphology and subglacial lakes is an indirect one since it is basal shear stress and ice dynamics that controls the ice-sheet surface profile and not solely the existence of subglacial water. For instance, an area of the ice-sheet base where no subglacial lake exists, but where very low shear stresses occur (e.g., a region of water-saturated unlithified sediments) may also have a nearly-flat surface region above it (Siegert and Ridley, 1998a).

Not all subglacial lakes have nearly-flat ice-sheet surface features associated with them. For example small lakes do not show this feature, and neither do lakes that are located very close to or at the ice divide (Siegert and Ridley, 1998a).

To date, the most extensive investigation into the flat regions above subglacial lakes analysed ERS-1 altimeter information across the Dome C region, where a large cluster of lakes are located (Siegert and Ridley, 1998a). Analysing ERS-1 satellite radar altimetric data from East Antarctica provided information on (1) the surface area of known subglacial lakes, identified from the calculated area of the nearly-flat surface feature (as in Kapitsa et al. (1996) for the Vostok lake), and (2) the minimum size of known subglacial lakes which cause the formation of a flat region within the ice-sheet surface. If flat ice-surface features are identified where RES data show the absence of a subglacial lake, these RES data may can be used to determine the appropriate, alternative ice-sheet basal conditions. This point is referred to later in a section on the geological significance of subglacial lakes.

Analysis of the satellite radar altimeter map, in conjunction with the locations of subglacial lakes beneath Dome C, shows that 20 of 36 RES lake reflectors lie beneath an ice-sheet surface slope of less than 0.01°. In general, relatively larger lakes occur beneath nearly-flat regions (Dowdeswell and Siegert, 1999). The mean minimum length of RES lake-type reflectors beneath 'flat' regions is ~8.3 km, whilst it is ~3.3 km if no surface expression is observed. Consequently, the occurrence or non-occurrence of flat ice-sheet surface regions is controlled by the size of the lake (Fig. 7). This has implications for the minimum size of subglacial lakes that can be detected from satellite information, and leads to a suggestion that there may be more subglacial lakes ($\leq$ 4 km in length) which lie undetected between the 50–100 km separations of the RES flight tracks (Dowdeswell and Siegert, 1999).

Measurements of lake extent from satellite data have implications for the calculated minimum length of lakes, determined previously from RES data. It can be assumed that the shape of the ERS-1-derived near-flat surface regions, if associated with lake-type RES reflectors, represents the shape of subglacial lakes (e.g., Kapitsa et al., 1996). In support of this assumption, where RES data are available, the edge of the subglacial lakes corresponds well with the margin of the flat surface region. Consequently, where the ERS-1-derived lake dimensions are larger than those detailed in RES data, the minimum lengths and approximate areas can be recorded directly from the ice-surface features (Figs. 6 and 7). However, it must be stressed that without confirmation of subglacial lake existence from RES data, a flat region on the ice-sheet surface does not necessarily correspond with a subglacial lake beneath.

After consideration of the surface morphology of the ice sheet the number of subglacial lakes that exist in the Dome C region above the 3000 m contour,

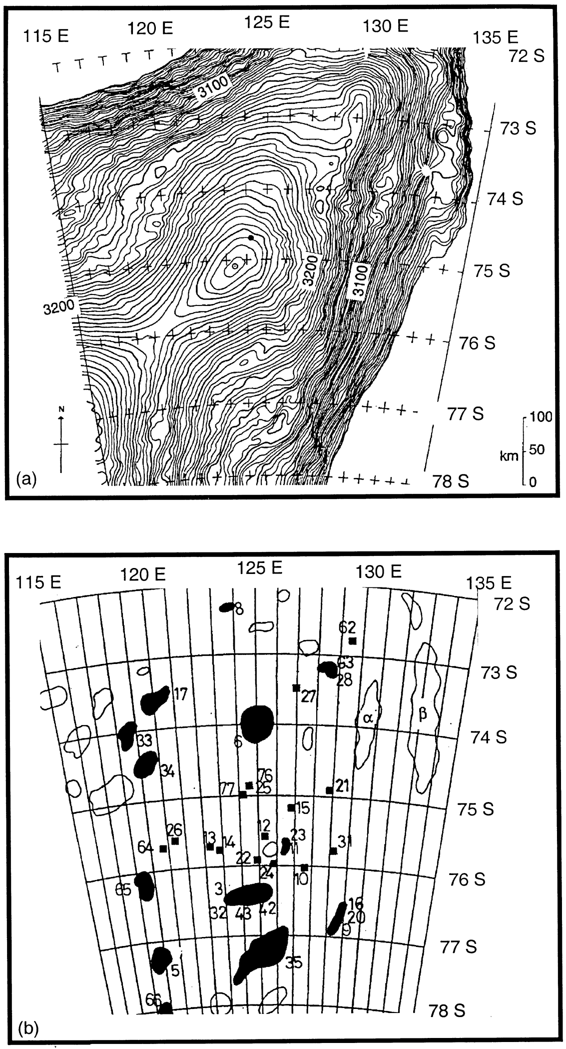


Fig. 6. (a) ERS-1 satellite radar altimeter data from the Dome C region of central East Antarctica. The location of subglacial lakes with respect to 'flat regions' on the ice surface are indicated. Contours are given in 2-m intervals. Dome C Station (filled circle) and Dome C Summit (open circle) are located. (b) The spatial coverage of flat surface regions determined by restricted areas of surface slope less than 0.01°. Those flat regions representing subglacial lakes (identified through analysis of RES data) are shown as black regions, whereas flat regions, beneath which RES data show no subglacial lake, are not filled. Taken from Siegert and Ridley (1998a).

was reduced from 36 to 28 (Siegert and Ridley, 1998a). The total surface extent of water masses at the base of the Dome C region was ~15,000 km^2. This value represents ~5% of the total basal area specified within the 3000 m satellite-derived contour (Fig. 6), and is compatible with the area of Lake Vostok.

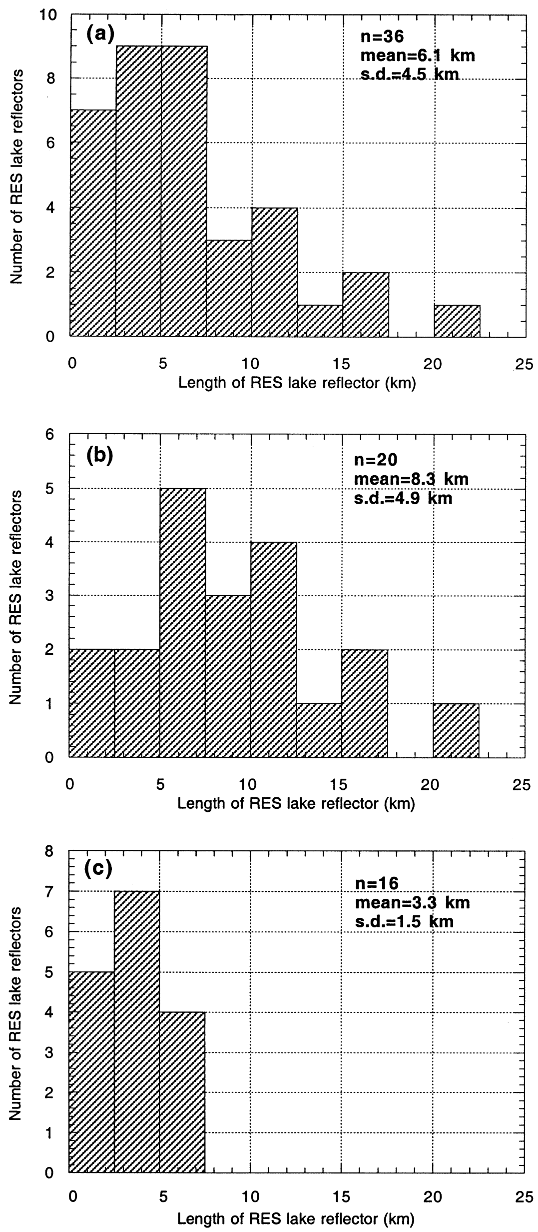


Fig. 7. Minimum length of RES subglacial lake records around Dome C. The measurements assume a constant aircraft-speed of 300 km h^{-1} (Siegert et al., 1996). (a) Total population of Dome C subglacial lake records above the 3000 m contour. (b) Dome C RES lake records originating from beneath flat regions in the ice-sheet surface. (c) Dome C RES records of subglacial lakes which do not exert a noticeable influence on the ice surface. The full population of the length records of known subglacial lakes is provided in Fig. 11.

3.2. Ice dynamics over subglacial lakes

As ice flows across the margin grounding line of a subglacial lake it experiences a rapid change in basal conditions, from grounded ice with considerable basal shear stress to ice afloat in water with negligible basal shear stress. Since Lake Vostok represents the most obvious example for studying ice flow across a subglacial lake, Siegert and Ridley (1998b) attempted to describe the velocity field in this region through consideration of: (1) ERS-1 satellite altimeter map; (2) internal structure of the ice sheet from RES information; and (3) consideration of the continuity of ice flux.

The only published horizontal ice velocity measured at Vostok Station has been given at 3 m year^{-1} in a 142°N direction, from astronomical investigations (Kapitsa et al., 1996). Assuming that ice velocity everywhere is controlled solely by ice surface slope, it would appear from a first inspection of the ice-sheet topography over Lake Vostok (Fig. 3), that the movement of the ice sheet changes direction from approximately eastward over the grounded regions, to southward over the lake. However, since the ice surface slope over the lake is very small compared with the surface gradient of the ice sheet around the lake, the ice velocity over the lake may also be influenced by the predominantly west–east flow of the surrounding grounded ice sheet. Unfortunately, without any models that describe basal sliding near the centre of ice sheets, or near the grounding lines (e.g., Hindmarsh, 1993), the relative importance of the surface slope compared with the surrounding grounded ice velocity, on the direction of ice flow over the lake cannot be calculated.

The surface of the ice sheet over Lake Vostok slopes predominantly from northwest to southeast over the northern region and from north to south over the southern area of the lake (with an angle of 0.004°), compared with the approximately west to east surface slope (0.08°) of the surrounding ice sheet resting over bedrock (Fig. 3). If it is assumed that the direction of ice flow over the sub-glacial lake is controlled solely by the surface slope (Fig. 3), then ice flow across the lake is generally west to east either side of the lake, and from NW to SE over the northern region, to N–S over the southern area, of the lake. However, since the conservation of ice flux must apply if the ice sheet is in relative steady-state, it must be concluded that the ice-sheet flow is more likely in an approximately W–E direction across the lake. The rationale behind this conclusion is as follows: (a) the ice-sheet surface slope over grounded ice to the west and east of the lake has a relatively constant value (0.08°), indicating a steady flux of ice from the western, and out of the eastern side, of the lake, (b) from this reasoning, a steady flux of ice occurs over the lake from west to east (i.e., approximately perpendicular to the direction of the low ice-sheet surface slope over the lake). However, under this rationale, the N–S slope of the ice sheet over the lake is not accounted for.

If ice flow were from north to south whilst over the lake then, from consideration of the continuity of ice mass, the velocity field in the central East Antarctic Ice Sheet would provide an unusually high flux of ice flowing southeastward beneath Vostok Station.

If one assumes N–S ice flow over the lake, then the direction of ice flow must change markedly across the western margin of the lake (i.e., from eastward to southward). In this situation, assuming that the ice-sheet is in steady-state, the flux of ice entering the 240 km long western margin of the lake must be similar to that across the downstream edge of the lake, at the 30 km-wide southern lake margin. Since the ice thickness only varies from 4200 m over the North, to 3700 m over the South of the lake, the velocity of ice across the southern margin of the lake would have to increase dramatically in order to provide ice flux continuity. Thus, the ice velocity beneath Vostok Station would have to be around eight times more than that over the western margin (assuming no ice is lost across the eastern side of the lake). The general velocity of ice in the central East Antarctic is up to ~ 3 m year^{-1} [a maximum value, estimated from numerical modelling studies of the region (Huybrechts, 1992)]. Thus, if ice flow is controlled solely by surface slope, the surface velocity of ice at Vostok Station would be up to ~ 24 m year^{-1}. Such an ice velocity is very high for the centre of a large ice sheet, and is inconsistent with the velocity measurement for this region (3 m year^{-1} in a 142°N). Moreover, there is no topographic expression in the surface elevation at Vostok, measured by the ERS-1 radar altimeter, for an unusually high velocity feature.

In order to account for both the N–S slope of ice, and the conservation of ice mass, it was proposed that the ice velocity field is dominated by the west–east movement of grounded ice, but that a N–S velocity component is established when the ice is located over the lake (Siegert and Ridley, 1998b). Under this scenario, the ice flowline turns clockwise and then anti-clockwise as ice moves from west to east across the lake (Fig. 4c). New interferometric SAR data used to calculate the ice velocity across Lake Vostok, is in agreement with this conceptual model of ice flow (R. Kwok, pers. comm.).

Thus, Lake Vostok exerts a significant influence on the dynamics of the ice sheet which lies over it. In addition, the nearly constant surface slopes to the west and east of the lake, suggest that Lake Vostok does not influence ice-sheet behaviour away from its margins.

To recap, the systematic analysis of the Antarctic RES database identified 68 sub-glacial lake-type reflectors, within several regions of Antarctica. Many of these subglacial lakes have a noticeable topographic expression on the ice-sheet surface. Therefore, several subglacial lakes in addition to Lake Vostok, possess a dynamic control on the ice sheet. For example, around the Dome C region of East Antarctica, the surface area occupied by subglacial lakes is estimated to represent $\sim 5\%$ of the ice-sheet base (Siegert and Ridley, 1998a). Therefore, it can be inferred from analysis of the data from Lake Vostok that other subglacial lakes may exert a dynamic influence on at least 5% of the interior East Antarctic Ice Sheet.

3.3. Salinity of lake water and the formation of Antarctic subglacial lakes

There have been two proposals for the formation of Antarctic subglacial lakes. The first is that the lakes have been formed as part of the relatively stable ice-sheet configuration (e.g., Shoemaker, 1991). The second is that they represent pre-glacial water bodies (i.e., terrestrial lakes or regions of sea-floor) that were subsequently overridden by glacial ice. The former theory dictates that subglacial melting, as occurs within the central regions of Antarctica today from geothermal heating, is the source of subglacial lake water. The water is routed somehow into bedrock troughs (such as in Lake Vostok) where it collects to form a lake. Under this rationale, the salinity of the lake water is likely to be similar to that of the basal layers of ice. From what is understood about the salinity of deep ice within Antarctic ice cores, this salinity value is likely to be extremely low. However, under the latter proposal, if subglacial lakes are relics from past terrestrial lakes or sea-floor regions, then the salinity could be much higher. Thus, identification of the salinity of subglacial lakes is fundamental to understanding their formation.

There are two means by which the salinity of subglacial lake water can be established. First is through consideration of how the ice sheet effectively 'floats' in the subglacial lake (Oswald and Robin, 1973; Kapitsa et al., 1996). The second is by analysing faint e/m reflections that have been observed from the floors of a few subglacial lakes (Gorman and Siegert, 1999).

RES data collected along the 230 km length of Lake Vostok shows information regarding the ice thickness (accurate to 1.5%). When this is plotted against the ERS-1-derived surface elevation, the data should lie along a straight line, the gradient of which is proportional to the ratio of the density of ice and water (Fig. 8). From the Vostok ice core, the density of ice at the base is likely to be around 0.913 g cm^{-3} (Kapitsa et al., 1996). The gradient of the best fit line (Fig. 8) corresponds with a salinity value of between 0.00 and 0.05‰. Also provided on Fig. 8 is a line that corresponds with seawater, at a salinity value of 35‰, that does not fit well with the data.

The manner in which e/m radiation is attenuated through water is related to the conductivity. In turn, conductivity of water is dependent on salinity of the water. By measuring the decay of radio-waves in water, from analysis of the power reflected from subglacial lake floors, the electrical conductivity can be determined (Gorman and Siegert, 1999). This procedure has shown the electrical conductivity of

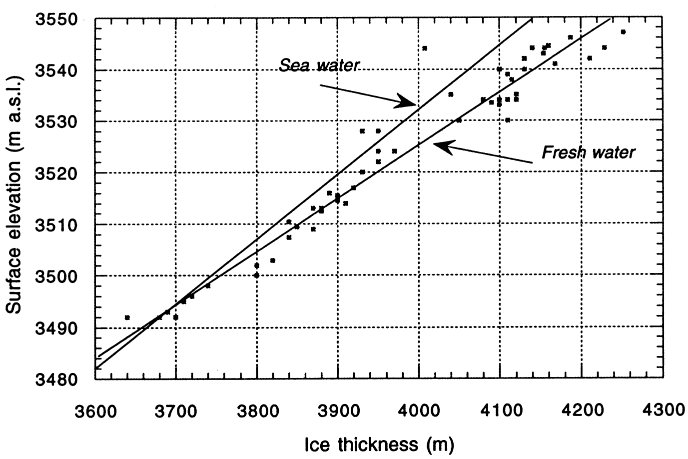


Fig. 8. The salinity of Antarctic subglacial lakes. Ice thickness vs. surface elevation for ice over Lake Vostok. Note how the gradient of the best fit line corresdponds with that for fresh water, whereas a seawater line does not link the data points. Taken from Kapitsa et al. (1996).

Antarctic subglacial lake water is between 3.08×10^{-4} mhos m^{-1} (for water-saturated sand at the lake floor) and 5.36×10^{-4} mhos m^{-1} (for bedrock). The electrical conductivity of fresh water is 10^{-4} mhos m^{-1}. Thus estimates of electrical conductivity within subglacial lakes show that the water is very pure and fresh (Gorman and Siegert, 1999).

4. Geological significance of Antarctic subglacial lakes

Antarctic subglacial lakes represent an important component of the central regions of glacierised Antarctica. Although the ice sheet may be relatively stable (i.e., it is neither in, nor about to enter, a period of rapid oscillations in volume and extent) the ice sheet exists as a dynamic configuration in which ice is constantly flowing towards the ice margin, and where the associated glacial erosion and sediment transport are active processes. Subglacial lakes have an influence on ice dynamics, act as sedimentary traps and house a vast amount of water. Under certain glaciological conditions subglacial lake water could be transported to the glacier margins. This is especially true for lakes located at the onset of enhanced ice flow. Subglacial lakes also allow us to determine thermal conditions within the ice sheet that are related to the geothermal heat flux through the Earth's crust. Thus, there are several areas of potential glacial–geological interest that can be explored through examination of Antarctic subglacial lakes.

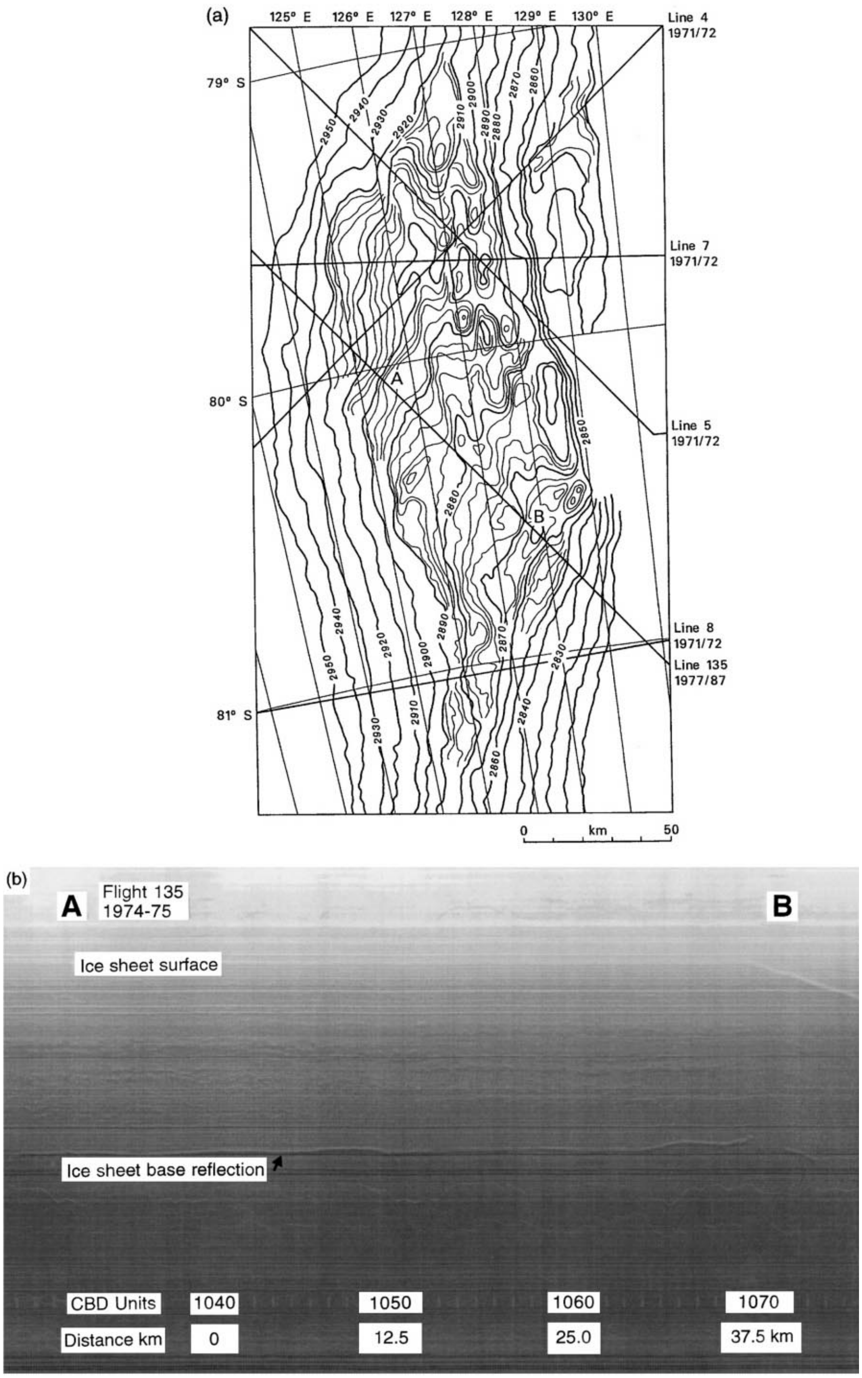
4.1. Sedimentation within subglacial lakes

The basal layers of ice within the Greenland Ice Sheet are characterised by the entrainment of sediment (as observed in the GRIP ice core). Although no ice core has reached the bed in East Antarctica, it seems relatively safe to assume that the base of East Antarctica will consist of sediment laden ice. Justification for this assumption is from observations of basal melting and sliding that will encourage the entrainment of sediment into ice (Hubbard and Sharp, 1993).

Zotikov (1987) identified a number of aspects involved in the process of sedimentation within subglacial lakes. The first is that ice flow across subglacial lakes occurs as a result of the decoupled ice-water interface. Thus, ice flow across the lake at the ice-sheet base will occur, resulting in a relatively constant supply of 'dirty' basal ice to the surface of the lake. Secondly, thermodynamic modelling of the ice sheet indicates that the rate of basal melting will be of the order of 1 mm $year^{-1}$. Because of this, the sedimentation rate is likely to be very low. However, subglacial lakes may be several million years old. Consequently, even under relatively minute sedimentation rates, a significant thickness of sediment would build up. Zotikov (1987) suggested that sediment build-up within subglacial lakes may be 10's to 100's of metres in thickness. In addition, the sedimentary record within subglacial lakes would demonstrate a complete history of Antarctic ice-sheet variability, dating back to the time at which the lake was formed (possibly several millions of years). No direct measurements of sedimentation within subglacial lakes have been made.

In addition to the flat ice-sheet surface regions identified at Dome C (Fig. 9), an inspection of the Antarctic ERS-1 satellite altimeter data has revealed a relatively large flat region which appears analogous to that in the Vostok area (Fig. 3a). This surface feature is located at 80°S, 127°E, and has a surface slope of 0.02° within a regional surface dip of 0.09° (Fig. 9). The surface slope is twice the value used in the criterion for flat region identification, but the sudden change in slope from that of the surrounding ice sheet (which is similar to that observed over the Vostok lake), means that it should be investigated as a potential surface feature resulting from the presence of a subglacial lake. The RES data for this region indicate a relatively undulating basal return (Fig. 9b) which suggests that a subglacial lake is not present in this area (reflections from subglacial lakes are unusually flat or mirror-like). However, the RES signal does show relatively bright subglacial returns which may be indicative of radio-wave reflections in the presence of water at the ice-sheet base. Siegert and Ridley (1998a) suggest that the flat region on the ice surface may be caused by a reduction in the basal shear stress (similar to that which occurs over subglacial lakes) due to the existence of geotechnically weak water-saturated basal sediment.

If water-saturated sediments are capable of inducing flat regions on the ice surface (appearing similar to those formed by subglacial lakes), then it can be concluded that the identification of subglacial lakes cannot be made through the interpretation of satellite radar altimeter data alone. However, satellite radar altimeter data can help to determine areas where subglacial lakes may exist, and where additional



RES data are required to confirm the presence of significant subglacial water bodies. Such features may represent subglacial lakes that have been filled with glacier-derived sediment.

4.2. Thermal regime above subglacial lakes

The existence of large volumes of water beneath the East Antarctic Ice Sheet has significant consequences for analysing the thermal regime of the ice sheet. The presence of sub-glacial lakes indicates that, in the region of the ice sheet over and adjacent to the lake, the temperature of the ice-sheet base is at the pressure melting point (Siegert and Dowdeswell, 1996).

Basal ice-sheet temperatures are controlled by a number of parameters including ice thickness, ice-sheet surface temperature and accumulation rate, heat transported through horizontal advection of ice, the basal heat gradient (the sum of geothermal heat flux, and heat produced from basal sliding) and heat derived from internal ice deformation. The basal heat gradient is dependent on the flow at the ice-sheet base, and its magnitude will therefore increase with distance from the ice divide as ice-sheet velocity increases. However, at the ice divide, the sole supply of basal heat will be from the geothermal processes. A simple numerical model of the ice-sheet thermal regime in such regions will indicate the geothermal energy required to allow basal melting to take place.

A model that can be used for such a purpose was developed by Robin (1955). The equation used is:

$$T_{\mathrm{B}} = T_{\mathrm{S}} - \frac{\sqrt{\pi}}{2} l \left(\frac{\mathrm{d}T}{\mathrm{d}z} \right)_{\mathrm{B}} \mathrm{erf}(h/l) \tag{1}$$

where

$$l = \sqrt{\frac{2kh}{b}} \tag{2}$$

and

$$\left(\frac{\mathrm{d}T}{\mathrm{d}z} \right)_{\mathrm{B}} = -\frac{\Lambda_{\mathrm{geo}}}{K} \tag{3}$$

Symbols used in the above equations are defined as follows: T_{B} is the basal temperature of the ice sheet (°C); T_{S} is the mean annual surface temperature of the ice sheet (°C); z is the coordinate in the vertical direction, positive upwards and zero at the ice-sheet base; h is the ice thickness above the sub-glacial lake (m); b is the mean annual surface accumulation of the ice sheet above the lake (m year^{-1}); k is the thermal diffusivity of ice (36.3 m^2 year^{-1}); K is the thermal conductivity of ice (2.10 W m^{-1}°C^{-1}); and Λ_{geo} is the Earth's geothermal heat flux (54 mW m^{-2}, unless stated otherwise). The Error Function (erf(h/l)) is a tabulated function just as a Sine or Logarithmic function (Abramowitz and Stegun, 1970).

As mentioned previously, the ice thickness, h, above a sub-glacial lake can be measured directly from the time-dependent raw radio echo sounding data. Assuming that the surface air temperature approximates the surface temperature of the ice sheet (Loewe, 1970), the surface temperature of the ice sheet, T_{S}, can be obtained from the map of mean annual air temperature, determined from field measurements (Robin, 1955). Surface accumulation on the ice-sheet surface above each lake, b, was obtained from a map of accumulation, interpolated from direct field measurements (Giovinetto and Bentley, 1985). Thus, assuming that the temperature above a subglacial lake is at the pressure melting value, Eq. (1) can be solved for Λ_{geo}.

Consideration of the spatial distribution of subglacial lakes shows that many ($\sim$ 40%) are located directly over, or relatively close (< 100 km) to, ice divides (Fig. 2). The ice sheet at and around ice divides will experience little basal heat derived from

Fig. 9. (a) ERS-1 satellite altimeter data from an unusually flat region to the south of Dome C. RES flightlines around this region are shown. Contours are given in 2 m intervals where the ice surface is relatively flat, and in 10 m intervals elsewhere. (b) Airborne RES information from flight no. 135, 1978–1979. The location markers AB refer to those indicated in (a). The basal reflection is identified as the non-flat line above the note 'ice-sheet base reflection'. The location of this flat surface region is provided in Fig. 1a. Note that the sub-parallel wavy lines in the lower half of (b) represent scratches on the original RES negative used to obtain the photograph. Taken from Siegert and Ridley (1998a).

horizontal ice motion terms. Consequently, the basal heat gradient required by Eq. (1) to calculate the pressure melting temperature above sub-glacial lakes located in such regions, will be associated mainly with the supply of geothermal heat. This geothermal heat gradient is calculated for sub-glacial lakes that lie along, or close to, the ice divide. In this experiment, the minimum geothermal heat flux (Λ_{min}) that allows the pressure melting temperature to be reached is determined. Variation of the calculated minimum basal heat gradient within the ice sheet close to ice divides will, therefore, be due mainly to changes in the geothermal heat flux within the Antarctic continent.

Comparing the geographical location of those sub-glacial lakes which exist near to ice divides with Λ_{min}, illustrates that the geothermal heat flux varies spatially around the Antarctic Plate (Fig. 10). For example, at Dome C, Talos Dome and Titan Dome, the datapoints illustrated in Fig. 10 are tightly clustered with little variation in the minimum geothermal heat flux (between 41 and 58 mW m^{-2}). However, the Λ_{min} around the large lake near to Vostok Station appears to remain at or below 43 mW m^{-2}, while at Ridge B it is between 37 and 42 mW m^{-2} (Fig. 10). For lakes located around the Hercules Dome region the minimum geothermal heat flux required for basal pressure melting is consistently above 60 mW m^{-2} (Fig. 10). Consequently, the geothermal heat flux in the Hercules Dome region of Antarctica may be around 20–25 mW m^{-2} higher than that in Ridge B, and 10-15 mW m^{-2} higher than in Dome C.

From the analysis of the geothermal heat properties beneath sub-glacial lakes, it can be suggested that, beneath ice divides of the East Antarctic Ice Sheet, the Antarctic Plate's geothermal heat flux varies between about 37 and 64 mW m^{-2}.

It should be noted that several datapoints in Fig. 10 relate to sub-glacial lakes that lie far (> 400 km) from an ice divide. In such cases, the minimum basal heat gradient (Eq. (3)) required to calculate the pressure melting value at the ice-sheet base should be regarded as a combination of the heat derived from the Earth plus horizontal ice-motion terms (i.e., internal ice deformation and basal sliding).

4.3. Subglacial water volume

Quantification of the exact volume of subglacial water held beneath the Antarctic Ice Sheet cannot be determined unequivocally. However, by evaluating (1) the area of subglacial lakes and maximum and minimum end-member values for (2) the water depth

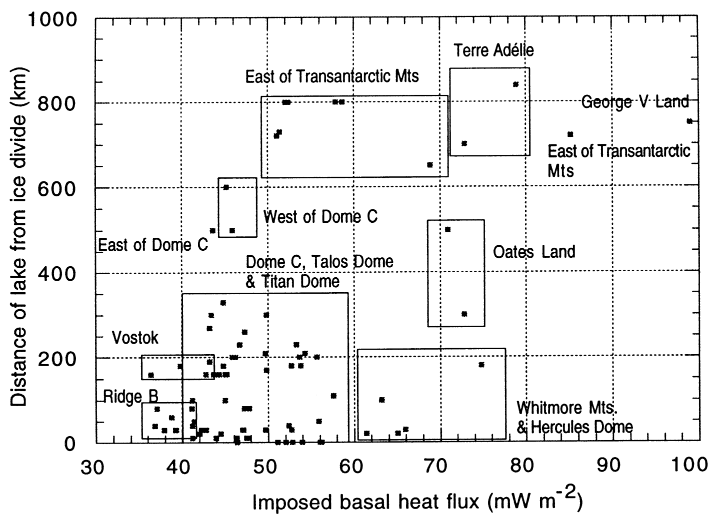


Fig. 10. Imposed basal heat flux required to induce pressure melting against the distance of the lake, along the direction of ice flow, from the nearest ice divide. Data points are identified with respect to geographical location. The graph illustrates the amount of basal heat required to ensure basal melting above sub-glacial lakes, and how this relates to the location of the lake within the ice sheet. See Fig. 2 for locations of named regions of Antarctica. Taken from Siegert and Dowdeswell (1996).

of subglacial lakes, based on consideration of RES and ERS-1 altimeter data, an envelope of water volumes can be established within which the actual basal water volume is likely to exist.

Taking RES-derived information on the total number of observed Antarctic sub-glacial lakes, their dimensions, and inferences from lake-marginal bed topography, order of magnitude calculations constraining the likely total water storage beneath the modern Antarctic Ice Sheet can be made. The likely magnitude of water storage in individual sub-glacial lakes, in addition to the volume estimate of ~2000 km^3 for Lake Vostok produced by Kapitsa et al. (1996) can also be estimated.

The surface area of 28 sub-glacial lakes (excluding Lake Vostok) calculated from analysis of ERS-1 satellite altimeter data is 15,000 km^2 (Siegert and Ridley, 1998a). If it is assumed that the remaining suite of Antarctic sub-glacial lakes identified from airborne RES are equi-dimensional, taking the lake-length frequency distribution in Fig. 11 yields a lake area of about 5000 km^2. Thus, the total lake area, excluding Lake Vostok, is around 20,000 km^2 (Dowdeswell and Siegert, 1999).

The RES dataset used to identify sub-glacial lakes coversed only 50% of the ice sheet (Drewry, 1983), leading to a second major source of underestimation of the total subglacial lake area for the Antarctic Ice Sheet. Assuming that sub-glacial water is generated primarily towards the centre of the ice sheet where ice is thickest (e.g., around Dome C), similar groups of sub-glacial lakes to those at Dome C and Ridge B may occur in comparably thick central regions of Antarctica (Dowdeswell and Siegert, 1999). For example, little RES work has been performed over the Bentley Sub-glacial Trench in West Antarctica (maximum ice thickness ~4000 m), and no flightlines were made over Valkyrjedomen (maximum ice thickness ~3500 m) and Dronning Maud Land (maximum ice thickness ~3500 m) in East Antarctica. Therefore, it appears likely that calculations of the area of Antarctic sub-glacial water bodies may be, as a minimum estimate, around one half of the total that is currently stored beneath the entire ice sheet (Dowdeswell and Siegert, 1999). Since the SPRI RES data cover only about half of the Antarctic Ice Sheet, the value of lake surface area can be doubled to approximately 40,000 km^2 for the whole ice sheet, excluding Lake Vostok (Dowdeswell and Siegert, 1999).

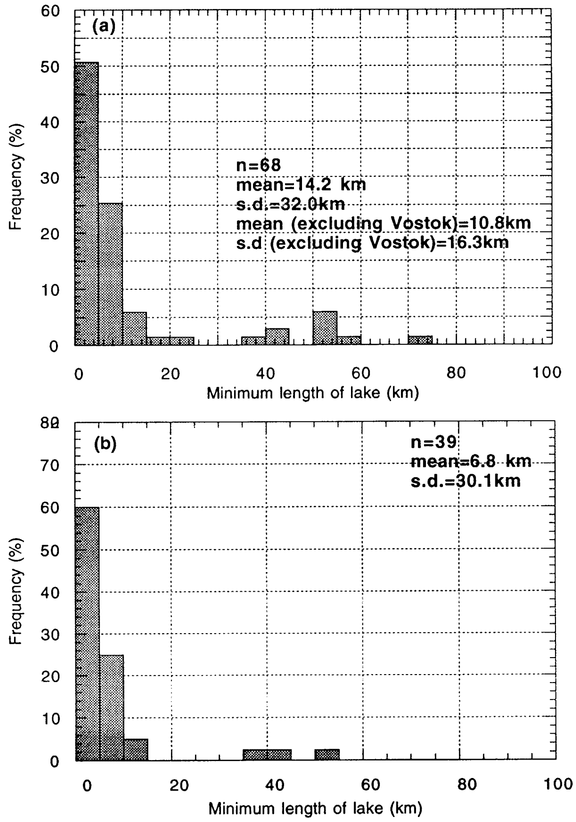


Fig. 11. (a) Frequency distribution of the length of observed sub-glacial lakes measured from radio-echo records. The mean value and standard deviation of 68 lake observations are given. Vostok Lake, at 230 km, is not shown, but is included in the accompanying statistics unless stated otherwise. (b) Frequency distribution for the length of 39 sub-glacial lakes that are not detected by ERS-1 altimeter. Taken from Dowdeswell and Siegert (1999).

Volume estimates for stored water beneath the ice sheet can be made by accounting for the water depth of subglacial lakes. Since radio-echo sounding provides only indicators of possible lake depth, based on measurements of the surrounding sub-glacial bedrock topography, a likely range of lake depths can be used to calculate an envelope of total sub-glacial water storage. First, using the rather high value of 500 m for average water depth in the lakes yields a total volume of 20,000 km^3. Dowdeswell and Siegert (1999) indicate that a more conservative series of mean water depths are 250 m, 100 m and 50 m, producing total stored water volumes of approximately 10,000 km^3, 4000 km^3 and 2000 km^3, respectively. To each of these values should be added the 2000 km^3 estimated by Kapitsa et al. (1996) for Lake Vostok. This then yields an envelope of between about 4000 and 12,000 km^3 for the likely volume of water stored in modern Antarctic sub-glacial lakes (Dowdeswell and Siegert, 1999).

4.4. Evacuation of subglacial water

The RES-derived measurements of both the total volume of water stored as lakes beneath the Antarctic Ice Sheet, and the size of individual water bodies, provides an analogue for conditions at the base of the large ice sheets which developed over mid and high northern latitudes during Quaternary full glacial conditions. The identification of subglacial lakes at the onset of enhanced ice flow is an important finding since these former ice sheets may have possessed a similar relationship between ice streams and basal water.

Several of Dowdeswell and Siegert's findings are of particular significance to: (a) the likely magnitude of outburst flood events from full glacial ice sheets and the release of waters to the global ocean, and (b) the consequences of such meltwater release for subglacial and proglacial landscapes, and for the marine geological record of, for example, Heinrich Events. Even if the water from all the observed Antarctic subglacial lakes was released at the same time, the volume involved would probably be less than 12,000 km^3. Given the very wide geographical distribution of lakes over the Antarctic Ice Sheet, it is highly unlikely that such synchronous drainage would take place. A water volume of 12,000 km^3 in Antarctic subglacial lakes is equivalent to about 35 mm of global sea level rise. If the largest subglacial lake, the 2000 km^3 lake at Vostok Station (Kapitsa et al., 1996), were to drain alone, a sea level equivalent of about 6 mm would be transferred to the global ocean. Other large lakes identified during RES investigations would, if drained to the ocean, probably produce less than 1000 km^3 of water or < 3 mm of sea level equivalent (Dowdeswell and Siegert, 1999).

It should be noted that the observations of Dowdeswell and Siegert (1999) are probably less representative of ice surface and basal conditions during Quaternary deglaciations, where large quantities of meltwater are likely to have been produced at the ablating surface and may have penetrated to the bed. However, the association between subglacial lakes and warm-based enhanced flow of ice may remain valid for these periods, since a hydrological connection between the lakes and ice margin is required for an outburst event.

5. Summary and conclusions

Airborne RES is capable of identifying subglacial lakes beneath up to 4 km of ice in Antarctica. Many subglacial lakes have a flat surface region on the ice sheet directly above them which can be surveyed by satellite altimetry. Using RES and ERS-1 altimeter information, the location and size of subglacial lakes can be established. A number of inventories of subglacial lakes have been compiled in the last 25 years. Oswald and Robin (1973) identified 17 subglacial lakes around Dome C and Ridge B. The large subglacial lake at Vostok Station was identified by Robin et al. (1977). An examination of post 1973 RES data by McIntyre (1983) increased this number to 64. However a re-examination of these data identified 77 subglacial lake-type reflectors (Siegert et al., 1996). The ERS-1 satellite altimeter can be used to determine the surface topography above subglacial lakes and, in so doing, provides information on which subglacial RES lake-type reflectors correspond with the same lakes. Siegert and Ridley (1998a) analysed these two datasets and concluded that 68 individual lakes are known to exist. This number can actually be increased to 71 after the identification of 3 further lakes in proximity with the Dome C station (Tabacco et al., 1998).

It is clear from Fig. 2 that the majority of the observed lakes are situated in relatively close proximity to ice divides, where both the surface slope and ice velocity are small. Two clusters of lakes, accounting for 70% of the total lake inventory, are in regions of Dome C and Ridge B in East Antarctica (Fig. 2). In addition, the ice divide stretching from West to East Antarctica (which runs close to the South Pole), has several sub-ice lakes along its length in the areas of Hercules Dome and Titan Dome (Fig. 2).

The largest subglacial lake exists beneath and up to 230 km to the north of Vostok Station, central

East Antarctica (Kapitsa et al., 1996). The extent of Lake Vostok (14,000 km^2) was able to be determined with accuracy due to the flat surface region that exists above the lake which is measurable from the ERS-1 altimeter (Fig. 3).

The sudden change in the surface slope of the ice sheet above Lake Vostok is due to a rapid alteration in the ice dynamics from grounded to floating ice, as ice flows over the margin of the lake. Assuming that the ice sheet is in relative steady-state, considerations of continuity of mass show that the ice flows over the lake in a direction controlled more by the predominantly eastward slope of the surrounding grounded ice sheet, than by the low southward slope of the ice sheet over the lake (Fig. 4c). The influence of Lake Vostok on ice dynamics is limited to its shorelines. Assuming this to be true for the 68 other sub-glacial lakes have been detected within Antarctica, Antarctic sub-glacial lakes may have a control on the dynamics of the central regions of the ice sheet, where they currently occupy $\sim$5% of the ice-sheet base.

The water depths of Antarctic lakes can be established from three methods. First is by seismic investigations, where the p-wave two-way travel-time difference between reflectors from the ice-water and the water-lake-floor interfaces yield allow the calculation of lake depth. The one depth-calculation has been made using the seismic technique at Lake Vostok recorded a water depth of 510 m a few km from Vostok Station. Second is by consideration of bedrock slopes at the margins of subglacial lakes. Calculations of these slopes from RES information show that, by extrapolating these slopes beneath lake surfaces, a water depth of several 10's–100's of metres is likely. The third method of determining lake-depth comes from analysis of e/m reflectors from the water-lake floor interface. In total, six of these reflectors have been identified, and indicate water depths of around 10-20 m in all cases. Thus, it can be concluded that subglacial lakes are not thin (cm scale) layers of water, but are 10's of metres deep in most cases.

Assuming the set of Antarctic sub-glacial lakes identified from airborne RES are equi-dimensional, the lake length–frequency distribution in Fig. 11a yields a total lake area of about 20,000 km^2, plus 14,000 km^2 for Lake Vostok. Given that SPRI RES data cover only about 50% of the ice sheet, this value is doubled to about 40,000 km^2.

An envelope of mean water depths of 250 m, 100 m and 50 m, produces total stored water volumes of approximately 10,000 km^3, 4000 km^3 and 2000 km^3, respectively. Adding the $\sim$2000 km^3 estimated by Kapitsa et al. (1996) for Lake Vostok, this gives an envelope of between about 4000 and 12,000 km^3 for the likely volume of water stored in contemporary Antarctic sub-glacial lakes. This represents a global sea-level equivalent of 10–35 mm.

From analysis of radio-wave attenuation in subglacial water, the conductivity of Antarctic sub-glacial water bodies varies between about 1-2 $\times 10^{-2}$ mhos m^{-1}; reflecting very fresh water (1–3 PSU) beneath the Antarctic Ice Sheet. This finding has implications for theories about the formation of subglacial lakes. There are two main ideas, the first of which is that subglacial lakes form from basal meltwater that collects within subglacial topographic hollows, the second is that lakes represent ancient lakes or sea bed which was overridden by ice during the onset of Antarctic glaciation.

RES data from the ice base beneath two flat surface regions within the Dome C district (identified as α and β in Fig. 6), and a further large flat region located at 80°S, 127°E (Fig. 9), do not indicate the existence of subglacial lakes. However, the relatively high strength of these RES signals suggest the existence of basal water and, conceivably, water-saturated basal sediments. The low yield stress of such material would influence the ice-sheet dynamics in a manner similar to the negligible basal stress above subglacial lakes. Consequently, the ice surface morphology above regions of deforming sediment may be comparable with surface features above subglacial lakes. In addition, the absence of subglacial lakes within deep troughs where basal melting is likely to be occurring, such as the Adventure Subglacial Trough, suggests that the transport of water must be occurring at the base of the central Antarctic Ice Sheet.

By assuming that the ice temperature about above subglacial lakes is equal to the pressure melting value, for the case of lakes that lie at or very close to the ice divide a simple thermal model allows a prediction of the geothermal heat flux at the base of the Antarctic Ice Sheet. The geothermal heat flux calculated to ensure basal melting of the ice sheet was between 37 and 64 mW m^{-2} (Fig. 10).

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