Antarctic Subglacial Lake Discharges

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Antarctic subglacial lakes were long time supposed to be relatively closed and stable environments with long residence times and slow circulations. This view has recently been challenged with evidence of active subglacial lake discharge underneath the Antarctic ice sheet. Satellite altimetry observations witnessed rapid changes in surface elevation across subglacial lakes over periods ranging from several months to more than a year, which were interpreted as subglacial lake discharge and subsequent lake filling, and which seem to be a common and widespread feature. Such discharges are comparable to jökulhlaups and can be modeled that way using the Nye-Röthlisberger theory. Considering the ice at the base of the ice sheet at pressure melting point, subglacial conduits are sustainable over periods of more than a year and over distances of several hundreds of kilometers. Coupling of an ice sheet model to a subglacial lake system demonstrated that small changes in surface slope are sufficient to start and sustain episodic subglacial drainage events on decadal time scales. Therefore, lake discharge may well be a common feature of the subglacial hydrological system, influencing the behavior of large ice sheets, especially when subglacial lakes are perched at or near the onset of large outlet glaciers and ice streams. While most of the observed discharge events are relatively small (10^1–10^2 m^3 s^{-1}), evidence for larger subglacial discharges is found in ice free areas bordering Antarctica, and witnessing subglacial floods of more than 10^6 m^3 s^{-1} that occurred during the middle Miocene.

How still,
How strangely still
The water is today,
It is not good
For water
To be so still that way.

Hughes [2001, p. 48]

1. INTRODUCTION

Subglacial lakes are omnipresent underneath the Antarctic ice sheet. A recent inventory [*Smith et al.*, 2009a] brings the total to more than 270 (Figure 1), i.e., 145 from an inventory by *Siegert et al.* [2005] and more than 130 added since [*Bell et al.*, 2006, 2007; *Carter et al.*, 2007; *Popov and Masolov*, 2007; *Smith et al.*, 2009a]. Subglacial lakes are usually identified from radio echo sounding and characterized by a strong basal reflector and a constant echo strength (corroborating a smooth surface). They are brighter than their surroundings by at least 2 dB (relatively bright) and both are consistently reflective (specular). They are therefore called "definite" lakes [*Carter et al.*, 2007]. The larger ones are characterized by a flat surface compared to the surroundings with a slope around one tenth, and in the opposite direction to, the lake surface slope [*Siegert et al.*, 2005], hence the ice column above such a subglacial lake is in hydrostatic equilibrium. "Fuzzy" lakes are defined by high absolute and relative reflection coefficients, but are not specular [*Carter et al.*,

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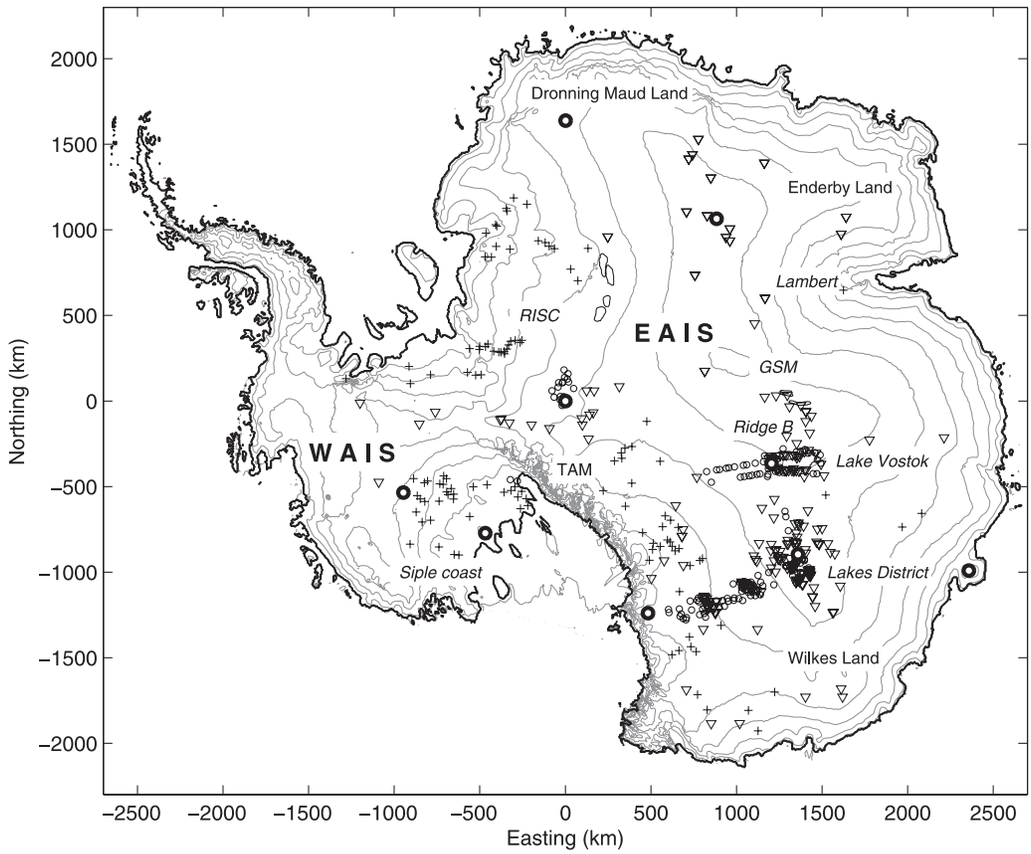

Figure 1. Surface topography [*Bamber et al.*, 2009] of the Antarctic ice sheet with radio echo sounding-identified large subglacial lakes (black lines), "definite" (inverted triangles) subglacial lakes [*Siegert et al.*, 2005; *Popov and Masolov*, 2007; *Carter et al.*, 2007], "fuzzy" (small circles) lakes [*Carter et al.*, 2007] and "active" (plus signs) lakes [*Smith et al.*, 2009a]. The bold circles show the position of deep ice core drill sites. Abbreviations are WAIS, West Antarctic Ice Sheet; EAIS, East Antarctic Ice Sheet; TAM, Transantarctic Mountains; RISC, Recovery Ice Stream Catchment; GSM, Gamburtsev Subglacial Mountains.

2007]. They are interpreted as corresponding to saturated basal sediments [*Peters et al.*, 2005; *Carter et al.*, 2007]. Most lakes lie under a thick ice cover of >3500 m and are therefore situated close to ice divides. Although the majority of subglacial lakes are small (<20 km in length), the largest lake is Subglacial Lake Vostok, containing 5400 ± 1600 km^3 of water [*Studinger et al.*, 2004]. Assuming an average water depth of 1000 m for large lakes and 100 m for shallow lakes, the volume of water in known Antarctic subglacial lakes is ~22,000 km^3, or approximately 25% of the water worldwide in surface lakes [*SALE Workshop Report*, 2007]. This is equivalent to a uniform sheet of water ~1-m thick if spread out underneath the whole Antarctic ice sheet.

Subglacial lakes were long time supposed to be relatively closed and stable environments with long residence times and slow circulations [*Siegert et al.*, 2001; *Kapitsa et al.*, 1996; *Bell et al.*, 2002]. As long as the huge amount of water they contain is not moving around through a hydrological network, they pose no threat to the stability of the ice sheet. However, the idea of closed subglacial environments has recently been challenged with evidence of subglacial lake discharge. First, from ERS laser altimetry data and later on from NASA's Ice, Cloud and land Elevation Satellite (ICESat) laser altimetry data, more than 120 "active" subglacial lakes have been determined [*Wingham et al.*, 2006a; *Fricker et al.*, 2007; *Smith et al.*, 2009a]. These are spots where a rapid lowering or rising of the ice surface is detected, a phenomenon that has been interpreted as either a sudden drainage (over a period of several months) or a rapid lake infill with subglacial water drained from an upstream subglacial lake [*Fricker et al.*, 2007; *Fricker and Scambos*, 2009].

In this chapter, we will give an overview of subglacial lake drainage, how it is observed, what the possible mechanisms are and whether such events may influence the stability of the ice sheet.

2. OBSERVATIONS OF SUBGLACIAL LAKE DISCHARGE

2.1. Jökulhlaups

Subglacial lake drainage is probably the most common feature in active volcanic regions, where they are called jökulhlaups. They originally refer to the well-known subglacial outburst floods from Vatnajökull, Iceland, and are often triggered by volcanic subglacial eruptions. More generally, jökulhlaups describe any large and abrupt release of water from a subglacial or supraglacial lake. Well-documented jökulhlaups are those from the subglacial caldera Grímsvötn, beneath Vatnajökull [*Björnsson*, 2002]: high rates of geothermal heat flux cause enhanced subglacial melting, and subglacial hydraulic gradients direct this meltwater to Grímsvötn where it collects. Jökulhlaups occur when the lake level reaches within tens of meters of the hydraulic seal, from where water is released and flows subglacially to the ice cap margin [*Björnsson*, 1988]. During this phase, ice velocities may well increase up to threefold over an area up to 8 km wide around the subglacial flood path [*Magnusson et al.*, 2007]. Owing to the large quantity of subglacial meltwater produced by high geothermal heat rates, jökulhlaups lead to a significant impact on ice dynamics. Not surprisingly, this type of jökulhlaups is unlikely to occur in Antarctica due to low geothermal heat flux. Geothermal heat flow measurements in the Lakes District (East Antarctica) point to values as low as 40 mW m^{-2} [*Shapiro and Ritzwoller*, 2004], which is supported by both the temperature gradient in the Vostok ice core as by inverse modeling [*Siegert*, 2000]. Nevertheless, these low values are sufficient to guarantee bottom melting due to the insulating ice cover. Moreover, since large quantities of subglacial water are stored underneath the ice sheet (e.g., 5400 km^3 for Lake Vostok), there is a substantial potential for subglacial lake discharge.

Despite low geothermal heating, evidence from Antarctic jökulhlaups is at hand: a first documented outburst occurred near Casey Station, Law Dome, Antarctica [*Goodwin*, 1988]. The discharge event started in March 1985 and lasted 6 months, with occasional outbursts during the austral autumn and winter of 1986. An oxygen-isotope and solute analysis of the spilled water revealed that its origin was basal meltwater, originating from an ice-marginal subglacial lake. However, the extent and depth of the reservoir remained unknown.

2.2. Satellite Observations

With the advent of satellite image interferometry, it became possible to observe and monitor small vertical changes at the surface of the Earth. Such changes measured in the direction of the satellite can be mapped onto both a vertical and horizontal component. The technique revealed very powerful to map horizontal flow speeds of Antarctic and Greenland outlet glaciers and ice streams using a speckle-tracking technique [*Joughin*, 2002]. *Gray et al.* [2005] analyzed RADARSAT data from the 1997 Antarctic Mapping Mission and used them interferometrically to solve for the three-dimensional surface ice motion in the interior of the West Antarctic Ice Sheet. They found an area of ~125 km^2 in a tributary of the Kamb Ice Stream displaced vertically downward by up to 50 cm between 26 September and 18 October 1997. Similar upward and downward surface displacements were also noted in the Bindschadler Ice Stream. Both sites seemed to correspond to areas where basal water is apparently ponding (hydraulic potential well). The authors therefore suggested transient movement of pockets of subglacial water as the most likely cause for the vertical surface displacements.

A well-documented Antarctic subglacial lake discharge is reported in the vicinity of the Lakes District, central East Antarctica. *Wingham et al.* [2006a] observed ice sheet

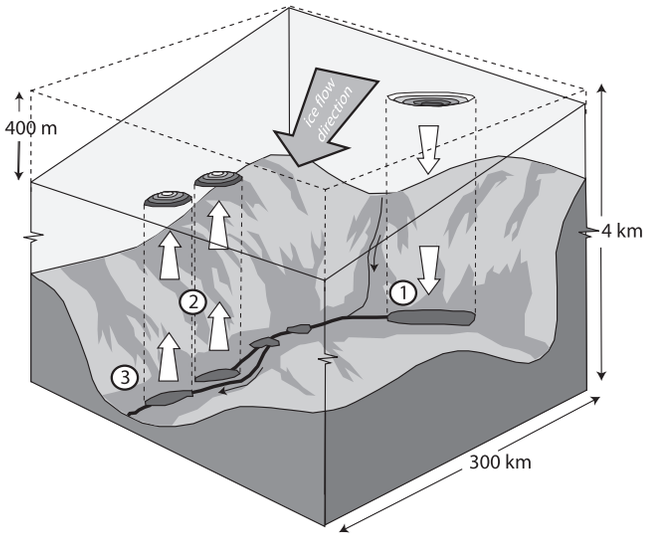

Figure 2. Rapid drainage and hydraulic connection of Antarctic subglacial lakes, inferred from satellite altimetric measurements of the ice sheet surface over the Adventure subglacial trench, East Antarctica between 1996 and 2003. Circled numbers indicate the following: 1, lake drainage results in ice sheet surface lowering of ~3 m over ~16 months; 2, ice sheet surface uplift occurs over a series of known subglacial lakes ~290 km from the initial lake drainage; 3, once surface lowering over the upstream lakes ceases (implying a reduction in the supply of water), ice surface lowering over at least one downstream lake occurs. Adapted from *Wingham et al.* [2006a].

surface elevation changes that were interpreted to represent rapid discharge from a subglacial lake. The altimeter survey by the satellite ERS-2 revealed two anomalies of ice sheet surface elevation change in the vicinity of the Adventure subglacial trench. One anomaly, at the northern (upstream) end of the trench, shows an abrupt fall in ice-surface elevation (Figure 2 and curve L1 in Figure 3). Some 290 km distant from L1, a corresponding abrupt rise occurred at the southern (downstream) end of the trench (Figure 2 and U1 and U2 in Figure 3). The only mechanism that explains these observations is a rapid transfer of basal water from a subglacial lake beneath the region of surface lowering to lakes beneath the regions of uplift. *Wingham et al.* [2006a] estimated the magnitude and rate of subglacial drainage of 1.8 km^3 of water transferred at a peak discharge of 50 $m^3\ s^{-1}$ during a period of 16 months.

High-resolution evidence of subglacial lake discharges was obtained with ICESat laser altimetry [*Fricker et al.*, 2007]. Satellite laser altimeter elevation profiles from 2003 to 2006 collected over the lower parts of Whillans and Mercer ice streams, West Antarctica, revealed 14 regions of temporally varying elevation, which were interpreted as the surface expression of subglacial water movement. Contrary to the Adventure trench lakes, perched near the ice divide of the East Antarctic ice sheet in a region of relatively slow ice movement, the lakes in West Antarctica were detected in an area of fast ice flow, close to the grounding line.

Subglacial Lake Engelhardt is one of the larger subglacial lakes in the area, situated just upstream of the grounding line and flanking Engelhardt Ice Ridge. Image differencing presented by *Fricker et al.* [2007] and *Fricker and Scambos* [2009] revealed a spatial extent of the drawdown of this lake of 339 km^2 ± 10%. This area is sufficiently large that the ice above the lake is in hydrostatic equilibrium, corroborated by the flatness of profiles across it (Figure 4). Draining and filling rates were estimated from the 2003–2006 flood and the total water loss over that period estimated as 2.0 km^3 [*Fricker and Scambos*, 2009], a value comparable to the Adventure trench discharge. Since Lake Engelhardt is situated ~7 km upstream from the grounding line, the floodwater is almost certainly discharged directly into the subglacial lake [*Fricker and Scambos*, 2009]. Since June 2006, Lake Engelhardt is steadily filling at a rate of 0.14 $km^3\ a^{-1}$ (Figure 4).

The analysis of *Fricker and Scambos* [2009] was further extended to the whole Antarctic ice sheet north of 86.6°S, based on 4.5 years (2003–2008) ICESat laser altimeter data [*Smith et al.*, 2009a]. This analysis detected 124 "active" lakes. Most of these lakes are situated in the coastal regions

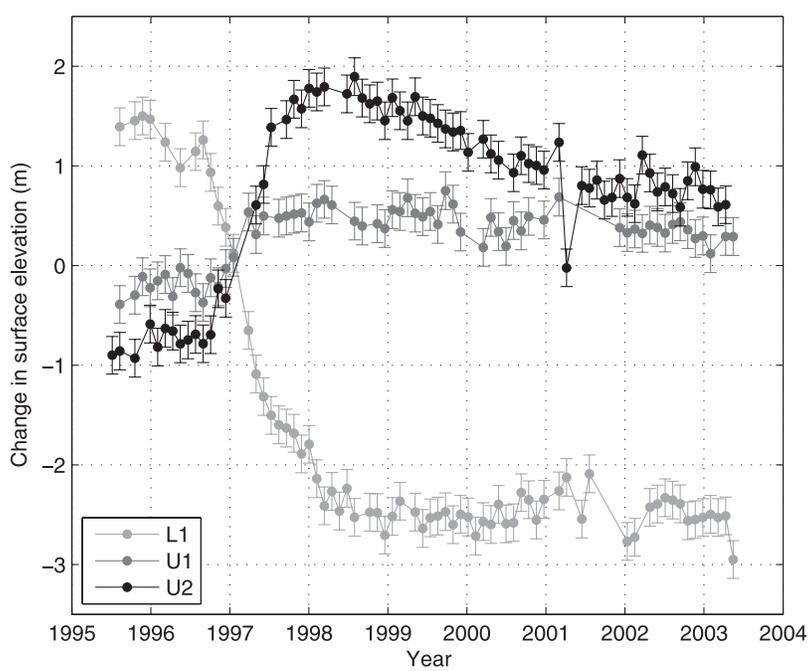

Figure 3. ERS-2 altimetric data from three sites, L1, U1, and U2, in the Adventure trench area (East Antarctica). Lake U3 has the same signature of U2 and is not shown here for clarity. The error bars are 1-sigma errors, which are determined empirically and determined as ±0.18 m. Adapted from *Wingham et al.* [2006a].

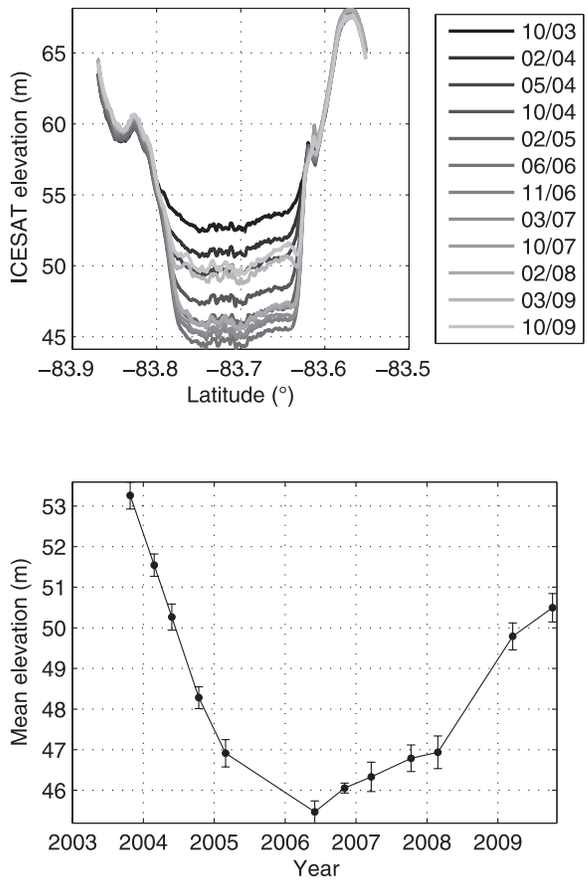

Figure 4. (top) Evolution of surface evolution across Engelhardt Lake (West Antarctica). (bottom) Mean elevation across the lake surface [*Fricker and Scambos*, 2009; H. A. Fricker, personal communication, 2009].

of the ice sheets and present underneath ice streams (Figure 1). They may, therefore, form reservoirs that may contribute pulses of water to produce rapid temporal changes in glacier speeds [*Smith et al.*, 2009a].

However, one should remain careful in interpreting rapid surface changes detected through satellite laser altimetry as sudden drainage of subglacial lakes. The interpretation of small-scale surface displacement to recover subglacial water movement is complicated by the fact that other subglacial processes can also result in surface deformation [*Smith et al.*, 2009a]. Ice-flow models show that local changes both in basal friction and in basal topography can produce changes in surface elevation [*Gudmundsson*, 2003; *Pattyn*, 2003, 2004; *Sergienko et al.*, 2007]. Furthermore, upward vertical changes may not always imply lake filling with water coming from upstream [*Pattyn*, 2008]. This point will be discussed later.

Besides satellite altimetry, present-day as well as past subglacial lake dynamics may be detected through other remote sensing techniques. Both melting and refreezing above subglacial lakes may be estimated by ice-penetrating radar by measuring along-flow changes in the thickness of the basal internal layers where thinning of the basal ice is indicative of melting and thickening points to freezing [*Bell et al.*, 2002; *Siegert et al.*, 2000; *Tikku et al.*, 2004]. Large distortions in basal internal layers may alternatively be indicative for subglacial lake discharge. *Siegert et al.* [2004] calculated from the convergence of a number of internal layers across the West Antarctic ice sheet that observed distortions would require basal melt rates of up to 6 cm a^{-1}, which are an order of magnitude greater than those calculated for the Siple Coast region. In view of the recent evidence regarding subglacial lake discharge in this area [*Fricker and Scambos*, 2009; *Smith et al.*, 2009a], large melting rates may witness such events as well.

2.3. Ice Flow Acceleration

Water plays a crucial role in ice sheet stability and the onset of ice streams. Since water may move between subglacial lakes by a rapid drainage, enhanced lubrication of ice streams and glaciers is expected, resulting in a speedup of ice flow. Many subglacial lakes sit at the onset of ice streams, hence have the potential to enhance the ice flow further downstream [*Siegert and Bamber*, 2000]. *Bell et al.* [2007] detected a number of large subglacial lakes at the onset region of the Recovery Glacier Ice Stream (RISC in Figure 1), where ice is moving at rates of 20 to 30 m a^{-1}. *Stearns et al.* [2008] report an observed acceleration of ice velocity on Byrd Glacier, East Antarctica, of about 10% of the original speed between December 2005 and February 2007. The acceleration extended along the entire 75-km glacier trunk, and its onset coincided with the discharge of about 1.7 km^3 of water from two large subglacial lakes located about 200 km upstream of the grounding line. Deceleration of the ice flow coincided with the termination of the flood. These findings provide direct evidence that an active lake drainage system can cause large and rapid changes in glacier dynamics. More spectacular interactions between floods and glacier surges are reported from Icelandic jökulhlaups [e.g., *Björnsson*, 1998].

Subglacial lakes may also play a crucial role in the redistribution of subglacial meltwater. Water piracy has recently been found a suitable mechanism in explaining the on and off switching of streaming flow, as is the case for the Siple Coast ice streams [*Anandakrishnan and Alley*, 1997] or Rutford Ice Stream [*Vaughan et al.*, 2008], a process that may be controlled by periodical subglacial lake discharge.

3. MECHANISMS OF SUBGLACIAL LAKE DISCHARGE

3.1. Hydraulic Geometry

Water is almost ubiquitously present underneath the Antarctic ice sheet, either stored in subglacial lakes or part of a subglacial drainage network that transports this water to the edge of the ice sheet. Free water can exist at the ice-bed contact and interstitially in subglacial sediment. The pressure p_w of subglacial water is an independent variable that varies temporally and spatially in a complicated manner that is determined by the balance between influx and outflux of water, the geometry of the subglacial water system, the physical properties of the glacier substrate, thermodynamic conditions near the ice-bed interface, and the ice overburden pressure [*Clarke*, 2005].

Glaciers are through buoyancy supported by the subglacial water pressure, and the magnitude of this support stems from comparing the water pressure p_w to the ice overburden pressure $p_i = \rho_i g h$, where ρ_i is the ice density, g the gravity acceleration, and H is the ice thickness. The effective pressure $p_e = p_i - p_w$ is a common measure for the importance of buoyancy [*Clarke*, 2005]. For subglacial lakes, where the overlying ice is in hydrostatic equilibrium with the underlying water column, it follows that $p_e = 0$. Subglacial water movement is controlled by gradients in the fluid potential, the latter defined by

$$\phi_w = p_w + \rho_w g z, \tag{1}$$

where ρ_w is the water density and z is the elevation ($z = z_b$ at the ice/bed contact, where z_b is the bed elevation). For a subglacial lake that obeys the condition $p_w \equiv p_i$, differentiating equation (1) leads to

$$\nabla\phi_w = \rho_i g \nabla z_s + (\rho_w - \rho_i) g \nabla z_b, \tag{2}$$

where ∇ is the gradient operator and z_s is the surface elevation [*Shreve*, 1972]. For $\rho_i = 910$ kg m^{-3} and $\rho_i = 1000$ kg m^{-3}, the first term on the right-hand side of equation (2) is roughly an order of magnitude larger than the second term. This means that subglacial water flow is mainly driven by the surface topography of the glacier and, to a lesser extent, by the bed topography. Water ponding in a subglacial lake implies that $\nabla\phi_w = 0$. This ponding threshold can therefore also be expressed as a simple relationship between surface and bed slopes, i.e.,

$$\frac{dz_s}{dx} = \frac{\rho_i - \rho_w}{\rho_i}\frac{dz_b}{dx}, \tag{3}$$

where x is the along-path distance coordinate. This implies that water can be pushed out of a subglacial cavity whenever the ice surface slope is larger than one tenth the adverse bed slope, or $dz_s/dx > -\frac{1}{10} dz_b/dx$.

Based on the hydraulic potential gradient, *Siegert et al.* [2007] and *Wright et al.* [2008] calculated subglacial water flow patterns. The steepest downslope gradient of the hydraulic potential indicates the direction of the water flow, from which flow paths can be calculated that predict the drainage network [*Wright et al.*, 2008]. An example of such a reconstruction is given in Figure 5, based on an initial distribution of local meltwater production using a thermomechanical ice sheet model with appropriate boundary conditions of surface temperature and geothermal heat flux, the latter corrected for the presence of subglacial lakes [*Pattyn*, 2010]. The input geometry is based on a 5-km resolution updated BEDMAP database [*Lythe and Vaughan*, 2001; *Pattyn*, 2010] and a resampled surface topography [*Bamber et al.*, 2009]. Figure 5 clearly shows that major drainage paths converge in the large outlet glaciers and ice streams, dominated by the surface topographic slopes.

Water transport is largely depending on the hydraulic geometry, whether water flows through an aquifer (or a sheet) or through a pipe. Numerous idealized drainage structures have been proposed to describe water flow at the base of a glacier, including ice-walled conduits [*Röthlisberger*, 1972; *Shreve*, 1972], bedrock channels [*Nye*, 1976], water films and sheets [*Weertman*, 1972; *Walder*, 1982; *Weertman and Birchfield*, 1983], linked cavities [*Walder*, 1986; *Kamb*, 1987], soft sediment canals [*Walder and Fowler*, 1994], or porous sediment sheets [*Clarke*, 1996]. For glaciers or ice sheets resting on a hard bed, there are essentially three possibilities for subglacial water flow, either through a system of channels, a linked cavity system, or drainage in a water film. Observed jökulhlaups generally drain through conduits that connect the subglacial lake to the edge of the ice cap. A similar theory can be applied to connecting Antarctic subglacial lakes.

Röthlisberger [1972] based his theory of water flow through a subglacial conduit on the Gauckler-Manning-Stricker formula for the mean velocity of turbulent flow, i.e.,

$$\mathbf{v}_w = \frac{Q}{S} = m_n^{-1} R^{\frac{2}{3}} \left(-\frac{1}{\rho_w g}\frac{\partial \phi_w}{\partial x} \right)^{\frac{1}{2}}, \tag{4}$$

where $m_n = 0.08$ m$^{-1/3}$s is the Manning coefficient determining the roughness of the conduit, Q is the water discharge, and R is the hydraulic radius, which is defined as the cross-sectional area over a wetted perimeter. For a conduit of semicircular cross-section, $S = \pi r^2/2$, or $r = \sqrt{2S/\pi}$, the wetted perimeter is defined as:

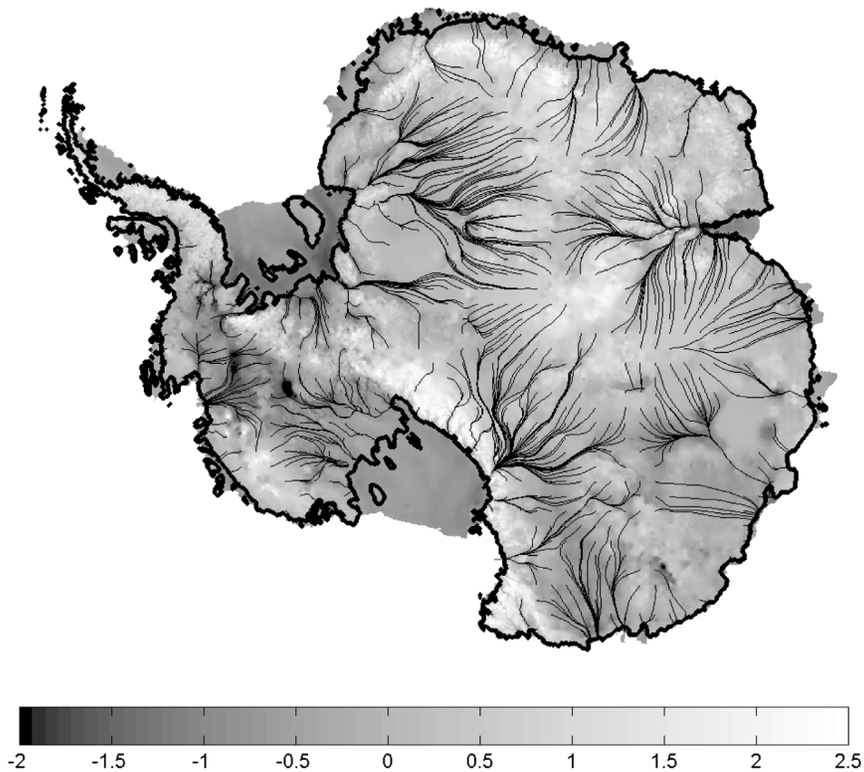

Figure 5. Calculated pattern of subglacial Antarctic drainage, superimposed on the subglacial topography (grey scale, in kilometers above sea level). Contrary to *Wright et al.* [2008], subglacial hollows where water is stored are not smoothed out, with the result that not all water drainage reaches the grounding line.

$$R = \frac{S}{\pi r + 2r} = \frac{S}{\pi\left(\frac{2S}{\pi}\right)^{\frac{1}{2}} + 2\left(\frac{2S}{\pi}\right)^{\frac{1}{2}}}. \tag{5}$$

Substituting equation (5) in equation (4) and rearranging terms, leads to an equation relating the size of the subglacial conduit to subglacial water discharge [*Röthlisberger*, 1972; *Peters et al.*, 2009]

$$S = \left[Q m_n \left(\frac{\sqrt{2}(\pi + 2)}{\sqrt{\pi}} \right)^{\frac{2}{3}} \left(-\frac{1}{\rho_w g} \frac{\partial \phi_w}{\partial x} \right)^{-\frac{1}{2}} \right]^{\frac{3}{4}}. \tag{6}$$

A similar and easier-to-derive expression can be found for a circular conduit, i.e.,

$$S = \left[Q m_n (2\sqrt{\pi})^{\frac{2}{3}} \left(-\frac{1}{\rho_w g} \frac{\partial \phi_w}{\partial x} \right)^{-\frac{1}{2}} \right]^{\frac{3}{4}}. \tag{7}$$

Peters et al. [2009] used equation (6) to derive the size of the conduit through which water would flow connecting two lakes along the hydraulic flow path in the Adventure trench basin, mentioned earlier. A similar approach was followed by *Wingham et al.* [2006a]. The geometry used is sketched in Figure 6: water from Lake L is drained via a conduit of l = 290 km in length over a period of 16 months. Owing to the discharge, the surface of the lake has lowered by Δh_L = 3 m. Since the lake size is estimated as S_L = 600 km^2 and assuming a cylindrical geometry, this leads to a water volume of V = 1.8 km^3 at a mean discharge of Q = 43 m^3 s^{-1}. The peak discharge of the lake has been estimated at 50 m^3 s^{-1} [*Wingham et al.*, 2006a]. Given a difference in ice thickness above both lakes of $\Delta H = -450$ m and the difference in height of the lakes (bedrock elevation difference) of Δz_b = 260 m, the hydraulic potential gradient can be approximated using equation (2), so that $\partial\phi_w/\partial x \approx -6.6$ Pa m^{-1}. Considering the system in equilibrium, the conduit size that sustains this discharge is obtained from equation (6), leading to a cross-sectional area of S = 78 m^2. This represents a semicircular conduit with a diameter of 14 m.

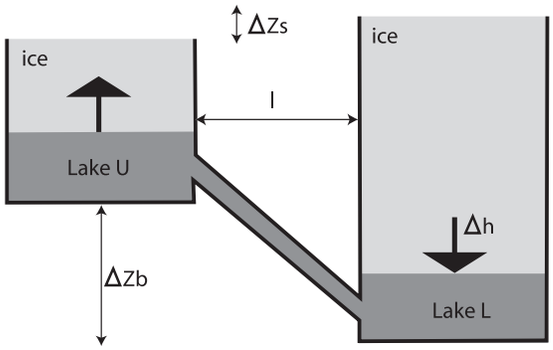

Figure 6. A simple model of the exchange between two lakes. An uphill flow of water is forced by the larger ice overburden at the lower Lake L. Both lakes are considered to have the same water depth. Adapted from the works of *Wingham et al.* [2006a] and *Peters et al.* [2009].

The question whether a conduit with that size can be formed between two lakes underneath the Antarctic ice sheet depends on the amount of energy that is available to form the conduit in the first place. Assuming again a cylindrical geometry for both lakes, the total energy release E_T is the energy due to the lowering of lake L (E_L) minus the energy needed to (1) raise lake U (E_U), (2) raise the water through the conduit by Δz_b (E_H), and (3) to maintain the water at pressure melting point (E_W). Assuming that the volume of water is conserved, i.e., $S_L \Delta h_L = S_U \Delta h_U$, it follows that

$$E_L - E_U = S_L \Delta h_L \rho_i g |\Delta H|, \tag{8}$$

$$E_H = V \rho_w g \Delta z_b, \tag{9}$$

$$E_W = -c_v V \frac{T_0}{L} \rho |\Delta H| \quad \left(\frac{1}{\rho_w} - \frac{1}{\rho_i}\right), \tag{10}$$

where T_0 = 273.15 K, c_v = 4.2 kJ kg^{-1} is the specific heat capacity of water, and $L = 3.3 \times 105$ J kg^{-1} is the latent heat for melting of ice. The total amount of energy equals $E_T = E_L - E_U - E_H - E_W$. Applied to the Adventure lake system, this leads to $E_T = 7.2 \times 10^{15} - 4.5 \times 10^{15} - 2.5 \times 10^{11} = 2.6 \times 10^{15}$ J. The energy required to melt a conduit of cross-sectional area S is $E_M = Sl\rho_i L$. As such, the available energy for melting a conduit over a distance l ($E_T = E_M$) is sufficient to sustain a conduit with a cross-sectional area S = 30 m^2 [*Peters et al.*, 2009]. This is about half the size of conduit that can be formed by the given discharge rate according to equation (6). However, in the former derivation, a relatively high Manning coefficient was used, typical for a rough conduit. Lower values of the Manning coefficient, e.g., 0.02 instead of 0.08 m$^{-1/3}$ s, corresponding to a smoother conduit, will give rise to a cross-sectional area of ≈30 m^2 [*Peters et al.*, 2009], as can be seen from Figure 7. Note also that the shape of the conduit (circular or semicircular) does not have a significant effect on the final conduit size corresponding to a given value for the Manning coefficient.

These calculations show that when considering the ice at the bed at pressure melting point, subglacial lake discharges can occur, and water can be transported over substantial distances of several hundreds of kilometers through a conduit with a diameter of 5 to 10 m.

3.2. The Nye-Röthlisberger Model

Rapid discharge of a subglacial lake is one of the different types of recognized jökulhlaups [*Roberts*, 2005]. The physical understanding of its time-dependent dynamics is based on empirical data from a rather limited number of events. Typical jökulhlaups from Grímsvötn (Iceland) increase toward a peak and fall rapidly. The periodicity is between 1 and 10 years, with peak discharges of 600 to 4–5 × 10^4 m^3 s^{-1} at the glacier margin, a duration of 2 days to 4 weeks, and a total water volume of 0.5–4.0 km^3 [*Björnsson*, 2002]. The difference with Antarctic subglacial lake discharge is striking: peak discharges are several orders of magnitude smaller and discharge duration generally more than a year. However, the periodicity may well be of the same order of magnitude, but the lack of longer time series prevents a proper evaluation.

The basic theory that governs this type of jökulhlaups is due to *Nye* [1976], extending the hydraulic theory of *Röthlisberger* [1972]. The theory assumes a single, straight subglacial conduit linking a meltwater reservoir directly to the terminus (or to another reservoir). For water to flow from one

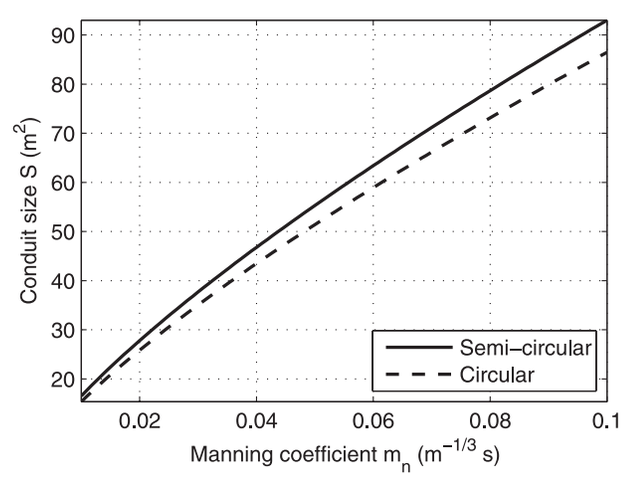

Figure 7. Conduit size S as a function of the Manning coefficient m_n for both a circular (7) and a semicircular (6) channel. Both curves are produced using the discharge geometry corresponding to the Adventure trench lakes.

to the other, a hydraulic gradient must exist. The size of the conduit will be due to a competition between the enlargement of the conduit and the processes that tend to shrink it. Energy is dissipated by flowing water, and some of the energy is transferred to the conduit walls causing ice to melt. Creep closure of the tunnel is due to ice deformation [*Clarke et al.*, 2004]. Nye derived partial differential equations describing the nonsteady flow in a conduit, accounting for the geometry and flow of ice, continuity of water, flow of water and energy, and heat transfer [*Nye*, 1976; *Fowler*, 2009]. *Peters et al.* [2009] applied the Nye-Röthlisberger model in its simplified form to the Adventure trench lakes, for which a hydrograph proxy exists (Figure 3).

Contrary to Icelandic jökulhlaups, temperatures of the lake water and the ice are likely to be similar at or close to pressure melting point, so that the energy equation and heat transfer equation can be simplified [*Peters et al.*, 2009]. The equations for the geometry and flow of ice, continuity and flow of water are

$$\frac{\partial S}{\partial t} = \frac{m_r}{\rho_i} - KS(p_i - p_w)^n, \tag{11}$$

$$\frac{\partial S}{\partial t} + \frac{\partial Q}{\partial x} = \frac{m_r}{\rho_w}, \tag{12}$$

$$\rho_w g \nabla z_b - \frac{\partial p_w}{\partial x} = f \rho_w g \frac{Q|Q|}{S^{8/3}}, \tag{13}$$

where K is a creep closure coefficient due to the nonlinear flow law $\dot{\varepsilon} = A\tau^n$, where $\dot{\varepsilon}$ is the strain rate and τ is the stress. More specifically, $K = 2A/n^n$ [*Fowler*, 2009]. The term m_r/ρ_i is the volumetric melt of the side-walls of the conduit, and f is a friction factor. *Wingham et al.* [2006a] used equation (11) to determine whether the subglacial conduit can remain open for a sustained period (at least more than a year). A fourth equation for energy (equation (19) in *Nye* [1976]), can be greatly reduced by removing all temperature dependencies and keeping the internal energy constant [*Peters et al.*, 2009]. However, as shown by *Spring and Hutter* [1982], this may lead to an overestimation of the peak discharge. Assuming steady state conditions (constant conduit size in time), it is possible to determine the effective pressure ($p_e = p_i - p_w$) to balance growth of the tunnel through melting with closure through ice flow. Neglecting sensible heat advection, the mean melt rate of the conduit is defined by

$$m_r = \frac{Q}{\rho_i L}\left(-\frac{\partial \phi}{\partial x}\right), \tag{14}$$

so that, by making use of the previously determined parameters, the melt rate of the Adventure lake conduit is estimated as $m_r = 8.3 \times 10^{-7}$ m^2 s^{-1}. Using $A = 2.5 \times 10^{-24}$ Pa^{-3} s^{-1} (the value of the flow parameter at pressure melting point) and a conduit size of 30 m^2, and rearranging equation (11) leads to p_e = 690 kPa. This is a very large value compared to the change in pressure at lake L by the discharge event ($\Delta p_L = \rho_i g \Delta h_L$ = 27 kPa). This led *Wingham et al.* [2006a] to conclude that the water flow was not stopped by the closure of the tunnel and that the lake could well have been emptied.

Peters et al. [2009] modeled the time-dependent evolution of a two-lake system, solving equations (11)–(13). They found that the Manning coefficient influences the peak discharge, but also the time for the flood to initiate as well as the lifetime of the flood. Furthermore, comparison with a circular conduit showed that the latter results in a faster onset and higher peak discharge. Considering more than two lakes shows the limits of the Nye-Röthlisberger model. It results in multipeak discharges, which were not observed along the Adventure trench [*Peters et al.*, 2009]. This may be due to the fact that the drainage system is not along a single conduit, but forms a distributed system based on broad shallow canals, as advocated by *Carter et al.* [2009]. Their study of the Adventure trench system is based on satellite altimetry and radio echo sounding data along the whole flow path and shows that the volume release from the source lake L exceeded the volume of the other lakes by >1 km^3, implying water loss of the system. This downstream release continued until at least 2003, when nearly 75% of the initial water release had traveled downstream from the filling lakes. Such discharge would only be sustained effectively by a broad shallow water system. A distributed system is also consistent with the 3-month delay between water release at the source lake and water arrival at the destination lake [*Carter et al.*, 2009]. Neither a porous aquifer or thin water layer at the ice bed interface are capable of transporting even a fraction of the inferred discharge. However, recent work by *Creyts and Schoof* [2009] suggests that distributed water sheets can be stable to much greater depth than previously quantified, as the presence of protrusions that bridge the ice-bed gap can stabilize them.

However, little remains known about how vertical deformations at the base of an ice sheet are transmitted to the surface, which is what is measured in the first place. Factors that influence this are, for instance, (1) the extent of the basal deformation anomaly compared to the ice thickness, (2) the time over which such deformation occurs, and (3) ice viscosity. A number of these effects will be explored in the next section.

4. EFFECT OF SUBGLACIAL LAKES AND LAKE DISCHARGE ON ICE FLOW

Observations by *Wingham et al.* [2006a], *Fricker et al.* [2007], and *Stearns et al.* [2008] suggest the rapid transfer of subglacial lake water and periodically flushing of subglacial

lakes connected with other lakes that consequently fill through a hydrological network. These observations point to similar events of similar magnitude. However, they stem from rapid changes at the surface from which the drainage events are inferred. With the exception of the Law Dome jökulhlaup [*Goodwin*, 1988], direct observation of such subglacial events is lacking and, therefore, knowledge on the mechanisms that trigger them as well. A major assumption in the previous analysis is that surface elevation changes are directly related to changes in subglacial water volume. This would be correct if ice deformation would not occur. One way, however, to investigate the effect of ice sheet dynamics on sudden changes in subglacial lake volume is to use a numerical ice sheet model. While most models applied to large-scale flow of ice sheets neglect longitudinal components, this is not applicable on the scale of subglacial lakes. Ice flow across a subglacial lake experiences no (basal) friction at the ice/water interface [*Pattyn*, 2003; *Sergienko et al.*, 2007]. As such, the ice column behaves as an "embedded" or "captured" ice shelf within the grounded ice sheet [*Pattyn*, 2003; *Pattyn et al.*, 2004; *Erlingsson*, 2006]. This is why the ice sheet surface across large subglacial lakes, such as Subglacial Lake Vostok, is relatively flat and featureless, consistent with the surface of an ice shelf. Since ice shelf deformation is governed by longitudinal or membrane stresses, effects of stretching and compression should be taken into account.

4.1. Subglacial Lakes and Ice Dynamics

In general, numerical ice sheet models are based on balance laws of mass and momentum, extended with a constitutive equation. Solving the complete momentum balance leads to a so-called full Stokes model of ice flow [*Martin et al.*, 2003; *Zwinger et al.*, 2007; *Pattyn*, 2008]. Further simplifications can be made to this system of equations. A common approach is the higher-order approximation, where it is assumed that the full vertical stress is balanced by the hydrostatic pressure [*Blatter*, 1995; *Pattyn*, 2003]. Further approximations to the Stokes flow, still including longitudinal stress gradients, are governed by the shallow-shelf approximation [*MacAyeal*, 1989; *Sergienko et al.*, 2007]. Not only is hydrostatic equilibrium in the vertical applied, but all vertical dependence, such as vertical shearing, is omitted. The whole system is integrated over the vertical, so that the Stokes problem is simplified to the two plane directions [*Hindmarsh*, 2004]. However, its applicability is reduced to areas of low basal frictions, such as ice shelves (subglacial lakes) and ice streams, as shown by *Sergienko et al.* [2007].

The simplest way of mimicking subglacial lake presence in an ice sheet model is to introduce a slippery spot within an area of high(er) basal friction. For this purpose, consider a uniform slab of ice of 80 by 80 km in size and $H = 1600$ m thick, lying on gently sloping bed ($\alpha = 0.115°$). The basal boundary condition is written as $\tau_b - \mathbf{v}_b\beta^2$, where τ_b is the basal drag, $\mathbf{v}_b$ is the basal velocity vector and β^2 is a friction coefficient. For large β^2, $\mathbf{v}_b$ is small or zero (ice is frozen to the bedrock); for $\beta^2 = 0$, ice experiences no friction at the base (slippery spot) as is the case for an ice shelf. In this experiment, the basal friction coefficient β^2 is defined by a sine function ranging between 0 and 20 kPa a m^{-1} (Figure 8d). Periodic lateral boundary conditions were applied, and the

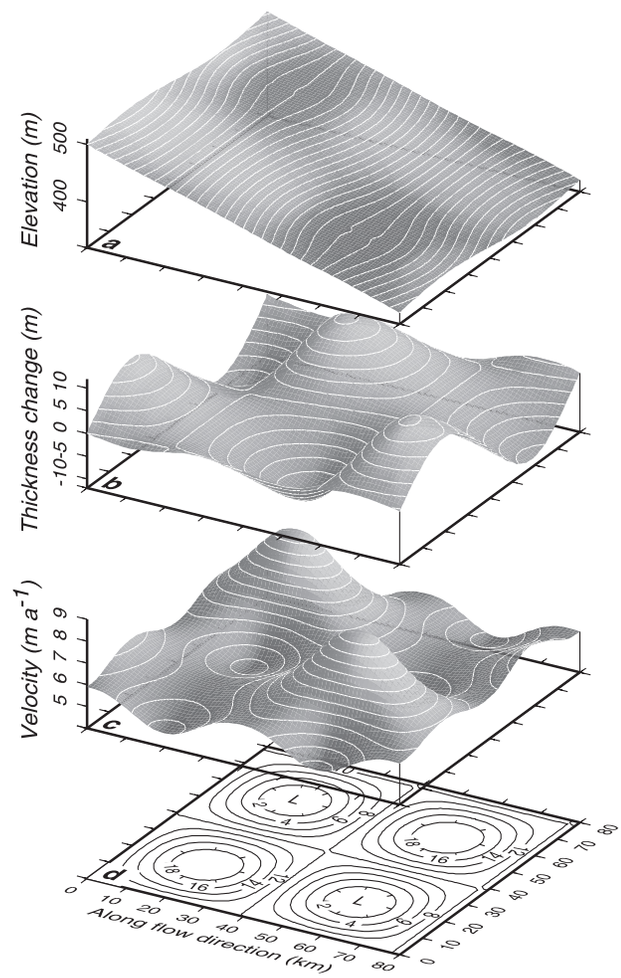

Figure 8. Effect of a slippery spot (subglacial lake) on ice sheet geometry and velocity field: (a) predicted steady state surface topography; (b) predicted change in ice thickness compared to initial uniform slab of 1600 m; (c) predicted horizontal surface velocity magnitude; (d) basal friction field β^2 varying between 0 (subglacial lake marked by L) and 20 kPa a m^{-1}. Ice flow is from left to right. From *Pattyn* [2004].

model was run to steady state. The effect of a slippery spot on the ice slab is shown by a local increase in ice velocity where friction is low (Figure 8c) as well as a flattening of the ice surface above this spot (Figure 8a). This flattening is due to a thinning of the ice upstream from the slippery spot and thickening of the ice downstream (Figure 8b) and is a direct result of the lack of basal shear across the slippery spot. *Gudmundsson* [2003] found a similar behavior for a linear viscous medium.

The effect of an embedded ice shelf is not only limited to a surface flattening, also the velocity field is influenced as the direction of the flow is not forced to follow the steepest surface gradient (as is the case when longitudinal stress gradients are neglected). This particularly applies to Subglacial Lake Vostok, where the observed surface flow is not along the steepest surface slope, since the latter is governed by the embedded ice shelf being in hydrostatic equilibrium with the underlying water body and the thickest ice, hence, highest surface elevation is found in the northern part of the lake. The effect of treating Lake Vostok as a slippery spot on the ice velocity is shown in Figure 9: the lack of a lake makes ice flow from north to south (right to left) or circumvents the lake, and no ice flow goes across Lake Vostok. Appropriate ice flow is obtained by treating Lake Vostok as a slippery spot, where ice flows from West to East (top to bottom), in agreement with observations [*Kwok et al.*, 2000; *Bell et al.*, 2002; *Tikku et al.*, 2004].

4.2. Effect of Lake Discharge on Ice Flow

As shown in Figures 3 and 4, subglacial lake discharge results in a sudden drop of surface elevation. The effect of such a sudden drainage can easily be mimicked by lowering the bedrock topography across a slippery spot that represents a subglacial lake. Using a numerical ice stream model applied to an idealized ice stream geometry, *Sergienko et al.* [2007] show that this effect significantly modulates the surface-elevation expression and that the observed surface elevation changes do not directly translate the basal elevation changes, due to the viscoplastic behavior of the ice when it flows across the lake. A sudden drop in ice surface will be filled in gradually due to the ice flow and its deformation. Therefore, subglacial water volume change is not directly proportional to the area integral of surface-elevation changes [*Sergienko et al.*, 2007].

In reality, the relation between subglacial lakes and the overlying ice sheet is even more complicated as the ice on top of larger subglacial lakes is in hydrostatic equilibrium. This is expressed by the fact that on most lakes, it holds that $\nabla z_s \approx -\frac{1}{10}\nabla z_b$. Whenever water is trapped in a subglacial cavity,

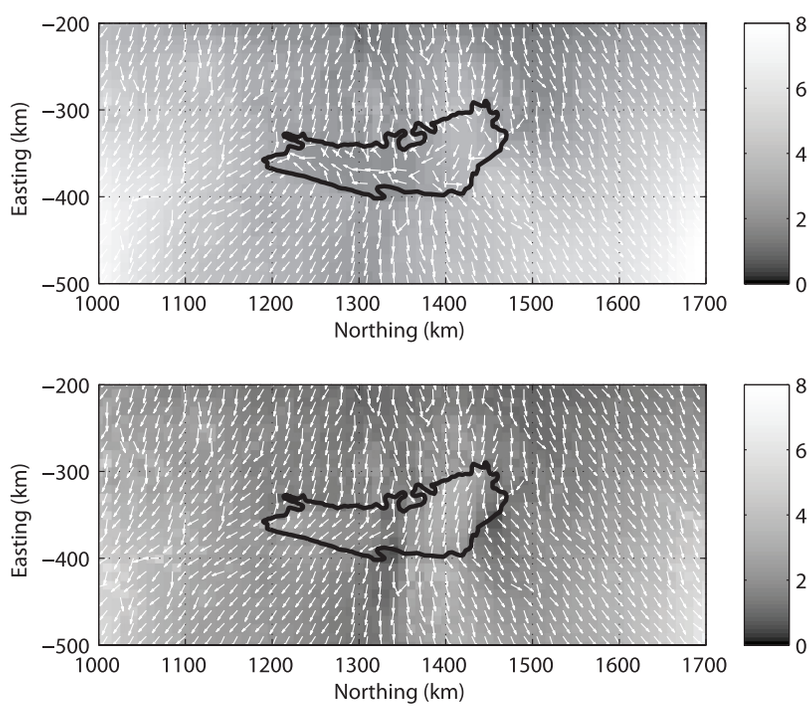

Figure 9. Predicted surface velocity (m a^{-1}) and flow direction across Lake Vostok for the "no lake" and "lake" experiment. The position of Lake Vostok is given by the black line. Adapted from *Pattyn et al.* [2004].

the effect will be transmitted to the surface by this relation. *Pattyn* [2008] implemented this effect in a full-Stokes ice sheet model of ice flow over a subglacial lake with a finite water volume. Full basal hydraulics, as those treated in the previous section, are not implemented in the model as such. A basal hydrological algorithm is used to check whether the lake "seal" breaks or not, based on the hydraulic potential gradient displayed in Figure 5. The hydrostatic seal is, thus, broken when water originating from the lake potentially arrives at the downstream edge of the domain. Nevertheless, as the lake empties, the marginal ice could collapse or fracture to form a cauldron, a common feature of jökulhlaups. The rate of subsidence is directly related to the rate of lake discharge, but requires an underpressure in the lake to draw down the overlying ice [*Evatt et al.*, 2006; *Evatt and Fowler*, 2007]. However, as small floods creating small surface depressions are considered here, only a flotation criterion for the ice over the lake is considered, as in the work of *Nye* [1976].

Conservation of water volume in the subglacial lake implies the definition of a buoyancy level (for a floating ice shelf this is sea level, but for a subglacial lake this is defined as $z_b + H\rho_i/\rho_w$). In an iterative procedure, the buoyancy level is determined for the local ice thickness to be in hydrostatic equilibrium on top of the lake for a given water volume. This procedure determines the position of both the upper and lower surfaces of the ice sheet across the subglacial lake in the Cartesian coordinate system. The contact surface between the ice sheet and the lake is then set to $\beta^2 = 0$.

The basic model setup is an idealized subglacial lake underneath an idealized ice sheet (Figure 10). The lake is defined as a Gaussian cavity. Ice thickness is $H_0 = 3500$ m, and the depth of the water cavity C_0 is taken as 400 m. Initially, the lake cavity was filled with 40×10^9 m^3 of water. The horizontal domain is L by L, where $L = 80$ km, and a grid size of 2 km was used (order of magnitude of ice thickness). The general characteristics of the steady state ice sheet geometry are typical for those of a slippery spot [*Pattyn*, 2003, 2004; *Pattyn et al.*, 2004], i.e., a flattened surface of the ice/air interface across the lake and the tilted lake ceiling in the opposite direction of the surface slope, due to hydrostatic equilibrium (Figure 10). The tilt of this surface in the direction of the ice flow will determine the stability of the lake, since the hydraulic gradient is dominated by surface slopes and therefore the flatter this air/ice surface the easier water is kept inside the lake cavity.

A perturbation experiment was carried out in which water is added to the lake at a rate of 0.1 m^3 s^{-1}. This corresponds to melting at the ice/lake ceiling at a rate of 2.5 mm a^{-1}, which is of the order of magnitude of basal melting to occur under ice sheets. It is regarded as a common gradual change in the glacial environment that on the time scales of decades

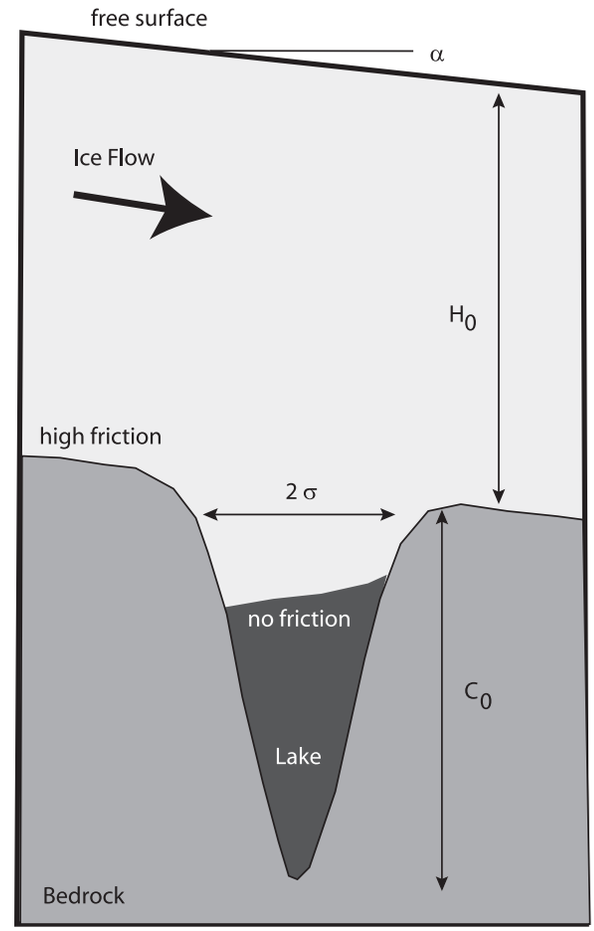

Figure 10. Model geometry of ice flow across an idealized subglacial lake. After *Pattyn* [2008]. Reprinted from the *Journal of Glaciology* with permission of the International Glaciological Society.

are hardly noticeable, hence, forming part of the natural variability of the system. Once the drainage condition is fulfilled, i.e., the hydrostatic seal is broken, the lake is drained at a rate of 50 m^3 s^{-1} for a period of 16 months, a value given by *Wingham et al.* [2006a] and which is of the same order of magnitude as the drainage rate of Lake Engelhardt [*Fricker et al.*, 2007]. This essentially means that enough energy is released to keep the subglacial tunnel open for a while and a siphon effect can take place. It is not clear whether this effect is only temporary or ends when the lake is completely emptied as suggested by *Wingham et al.* [2006a]. However, the drainage event of Lake Engelhardt suggests that water is still present in the lake after the event, as the surface aspect (flatness) has not changed over this period of time.

The time-dependent response of the lake system to such small perturbations is shown in Figure 11. When drainage

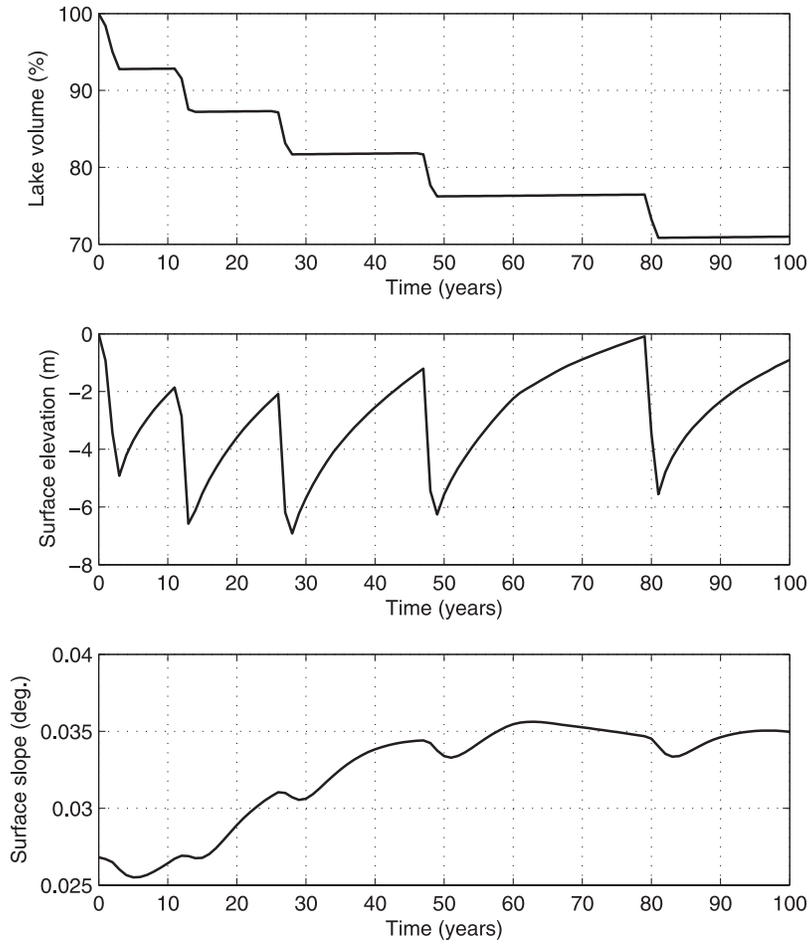

Figure 11. Time series of the perturbation experiment: (top) time evolution of lake water volume, (middle) ice surface elevation on top of the lake, and (bottom) ice surface slope across the lake. Adapted from *Pattyn* [2008].

occurs, episodic events take place, even though the initial geometric conditions are not met. The perturbation is, therefore, only a minor trigger to get the episodic lake drainage going. These events occur at a higher frequency at the beginning, i.e., when more water is present in the lake than later on and occur with frequencies of less than a decade. The episodic drainage results in variations in the surface elevation across the lake of rapid lowering, followed by a gradual increase until the initial level is more or less reached. This surface rise is clearly nonlinear, and the rate of uplift decreases with time. The frequency decreases with time as well as less water is present in the system, hence, hampering drainage. Interesting to note, however, is that the surface rise is not caused by an influx of subglacial water entering from upstream. On the contrary, less water is subsequently present in the whole subglacial system. Surface increase is due to an increased ice flux, filling up the surface depression created by the sudden lake drainage. Evidently, the maximum speed of surface increase is significantly lower than the surface lowering due to drainage.

As shown before, Nye's hydraulic theory offers an appropriate description of the drainage of subglacial lakes, especially where the lake is not connected to the ice surface, and the ice over the lake responds dynamically to lake level fluctuations [*Nye*, 1976]. According to that model, the magnitude and duration of floods appear to be controlled by the channel hydraulics. *Evatt et al.* [2006] show with a reduced hydraulic model based on Nye theory that the peak discharge is essentially related to lake volume, i.e., big lakes produce big floods. Therefore, small floods apparently observed by *Wingham et al.* [2006a] could be associated with drainage of small lakes or through channels at low effective pressure [*Evatt and Fowler*, 2007]. *Pattyn* [2008] supposed that once the hydrostatic seal is broken, a subglacial channel is established in which the water pressure is lower than the ice pressure. However, in order to keep the channel from closing

through viscous creep of ice, closure should be balanced by the melting of the channel walls. This energy is supplied by viscous dissipation of the turbulent water in the channel. As long as creep closure is balanced by the dissipated heat, the channel will remain open and discharge can take place. *Evatt et al.* [2006] have shown that lake drainage events as described by *Wingham et al.* [2006a] can occur over sustained periods of tens of months without the complete emptying of the lake. Furthermore, they also corroborate the fact that lake drainage is a common mechanism underneath the Antarctic ice sheet.

5. STABILITY OF SUBGLACIAL LAKES

The above-described mechanisms demonstrate that the stability of subglacial lakes is given by their nature of keeping the ice surface slope across the lake as small as possible due to the vanishing traction at the ice/water interface. The ice above Subglacial Lake Vostok, for instance, has a surface gradient of 0.0002 from north to south, which corresponds to 50 m of surface elevation change [*Siegert*, 2005]. If the slope of the grounded ice across the lake's western margin were changed, so too must the ice surface slope over the lake. Thus, the reason that a lake exists within the Vostok trough, and that ice-shelf flow is subsequently permitted, is due primarily to the flow direction of grounded ice upstream of the trough [*Siegert*, 2005]. Subglacial water will flow "uphill" if the ice surface slopes exceed 1/10 of the basal slope [*Shreve*, 1972]. Nevertheless, the ice surface gradient above Lake Vostok is 100 times less than the minimum basal slope of the head wall of the Vostok trough [*Studinger et al.*, 2003], which makes the lake stable unless the surface gradient increases tenfold, the value required to force the water out of the trough to the south [*Siegert and Ridley*, 1998]. However, the water level in Subglacial Lake Vostok, reaches close to a hydrostatic seal situated at the southeastern part of the lake, nearby Vostok Station [*Erlingsson*, 2006]. Water circulation modeling of Lake Vostok [*Thoma et al.*, 2007] points to an imbalance in the water mass balance of the lake which indicates either a constant growth of the lake or its continuous (or periodical) discharge into a subglacial drainage system. Even if periodic subglacial discharges occur of the order of magnitude as those observed by *Wingham et al.* [2006a], this would still remain unnoticeable for precise satellite altimetry due to the immense size of the lake.

Although the Antarctic ice sheet remained more or less stable and in its present configuration for at least the last 14 million years [*Kennett*, 1977; *Denton et al.*, 1993; *Huybrechts*, 1994], ice flow across the lake could also change drastically over less long time scales, such as glacial/interglacial cycles. These involve (1) changes in ice thickness and ice surface elevation and (2) migration of ice divides and alteration in the grounded flow direction. Surface elevation variations of the order of +50 m during interglacials and −100 m during glacials may well have occurred [*Ritz et al.*, 2001; *Siegert*, 2005]. Such changes may induce a migration of the ice divide, which traverses at present Lake Vostok from east to west [*Pattyn et al.*, 2004], thereby changing the ice thickness distribution. Since a change in ice thickness of >50 m at either side of the lake is sufficient to reverse the surface gradient across the lake, such an event will change the direction and magnitude of the ice/water slope. In certain cases, this may lead to a large unstable situation that potentially leads to a large drainage of Subglacial Lake Vostok.

6. EVIDENCE OF FORMER SUBGLACIAL OUTBURSTS

Evidence from catastrophic drainage of glacial and subglacial lakes stems from geomorphological evidence. The Channeled Scablands, for instance, cover an area of approximately 40,000 km^2 in Washington State (United States of America). They are the result of a catastrophic drainage of Glacial Lake Missoula, commonly known as the Missoula floods, at the end of the last glaciation. The floods are witnessed by a large anastomosis of flood channels and recessional gorges [*Baker et al.*, 1987]. *Clarke et al.* [1984] computed the water discharge from the ice-dammed lake through the ice dam, based on a modified Nye theory to account for effects of lake temperature and reservoir geometry [*Clarke*, 1982]. Maximum discharge was estimated as $2.7–13.7 \times 10^6$ m^3 s^{-1} [*Clarke et al.*, 1984].

Another major flood stems from the catastrophic discharge of Lake Agassiz, a proglacial lake formed at the southern margin of the Laurentide ice sheet during deglaciation of the latter and drained through a subglacial system [*Clarke et al.*, 2004]. Flood hydrographs for floods that originate in subglacial drainage conduits were simulated using the Spring-Hutter theory [*Clarke*, 2003], leading to a flood magnitude of $\sim 5 \times 10^6$ m^3 s^{-1} and a duration of half a year [*Clarke et al.*, 2004].

Spectacular meltwater features associated with subglacial outburst floods are also reported in the Transantarctic Mountains in southern Victoria Land, Antarctica. Since the Antarctic ice sheet has been relatively stable over the last 14 million years, they must predate Pleistocene glacial periods. Fortunately, and because of the long-term hyperarid polar climate, the outburst features are well preserved and suggest Miocene ice sheet overriding of the Transantarctic Mountains [*Denton and Sugden*, 2005]. These features consist of channels associated with areal scouring, scablands with scallops,

potholes, and plunge holes cut in sandstone and dolerite [*Denton and Sugden*, 2005]. Channels systems and canyons of as much as 600 m wide and 250 m deep can, for instance, be witnessed in the northern part of Wright Valley (Figure 12). Some of the observed potholes are >35 m deep [*Lewis et al.*, 2006]. These features are consistent with incision from subglacial meltwater, and the estimate discharge is of the order of $1.6–2.2 \times 10^6$ m^3 s^{-1} [*Lewis et al.*, 2006]. Their analysis also shows that the major channel incision predates 12.4 Ma, and that the last subglacial flood occurred sometime between 14.4 and 12.4 Ma ago. The source for such catastrophic discharge event should be a large subglacial lake. In view of the subglacial drainage flow paths (Figure 5) and the likelihood that the ice sheet was slightly bigger in size at that time, Subglacial Lake Vostok is a likely candidate. *Lewis et al.* [2006] advocate that the number and total volume of subglacial lakes beneath the East Antarctic ice sheet during the Middle Miocene could have been considerably greater than today due to the warmer basal conditions and larger ice sheet size. However, the present-day ice sheet has a predominant wet bed (reaching pressure-melting point) in the central parts where the ice is thickest, so that Miocene conditions could well be similar to present-day ones.

Besides the possible ice sheet instability due to the large quantity of subglacial water release, the huge freshwater discharge may also have an impact on deep-water formation in the Ross Embayment. Moreover, such type of subglacial floods could have formed a trigger for changes in middle Miocene climate [*Zachos et al.*, 2001; *Lewis et al.*, 2006].

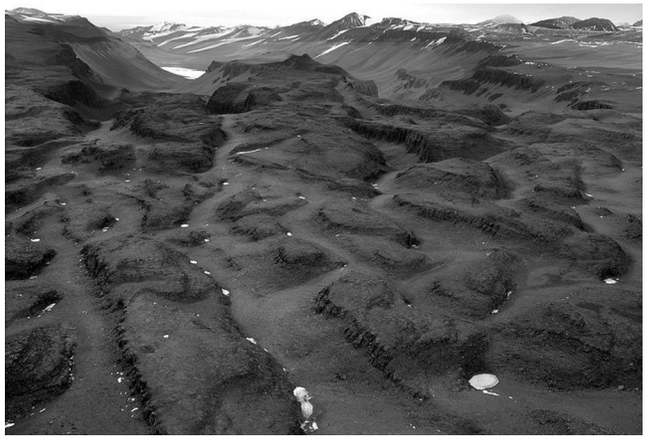

Figure 12. View of the Labyrinth, Wright Valley, Antarctica. The channelized system is formed by subglacial water routing due to a major Miocene subglacial drainage. From Photo Library, U.S. Antarctic Program. Photograph by P. Rejcek, National Science Foundation (2007).

Subglacial meltwater channel systems have also been detected offshore on the continental shelf of West Antarctica in the western Amundsen Sea Embayment [*Lowe and Anderson*, 2002, 2003; *Smith et al.*, 2009b]. They lie in the alignment of large outlet glaciers and ice streams, such as Pine Island and Thwaites Glaciers, both characterized by an important present-day dynamic ice loss [*Wingham et al.*, 2006b; *Shepherd and Wingham*, 2007; *Rignot et al.*, 2008]. The offshore meltwater systems relate to periods when the ice sheet was expanded and/or exhibiting a different dynamic behavior. While *Lewis et al.* [2006] investigate the morphology of subglacial channel systems, *Smith et al.* [2009b] focus on the sediment infill of the channels, since they have the potential to reveal important information on channel genesis and drainage processes. The presence of deformed till, for instance, at one core site and the absence of typical meltwater deposits (such as sorted sands and gravels) at other cores suggest that the channel incision predates the overriding by fast flowing ice streams during the last glacial period [*Smith et al.*, 2009b]. The channels were, therefore, likely formed over multiple glaciations, possible since Miocene as well [*Smith et al.*, 2009b].

7. CONCLUSIONS AND OUTLOOK

Recent observations on rapid discharge of water from subglacial lakes, as well as modeling of subglacial water drainage and interactions with the ice sheet, confirm that rapid subglacial discharge is a common feature of the Antarctic ice sheet. Although discharge rates seem rather small compared to observed jökulhlaups of Vatnajökull, evidence of former catastrophic outbursts exists, and they probably predate the Pleistocene epoch. These events are of a similar order of magnitude as those witnessed in North America during the deglaciation of the Laurentide ice sheet (e.g., Missoula floods).

Theory of subglacial lake drainage, developed several decades ago, has been improved over recent years and has been applied to Antarctic subglacial lake discharges. It shows that sustained drainage between linked lakes over distances of several hundreds of kilometers is possible for conduits of tens of meters in diameter. Linking discharge events with viscoplastic ice sheet models exhibits a nonlinear response of the ice surface to sudden discharge events, hence, complicating the interpretation of satellite altimetry observations.

Future challenges therefore lie in a better understanding of the dynamics of Antarctic subglacial hydrology, its relation to overall ice sheet dynamics, and its possible influence on the general climate evolution. Evidence presented here supports the idea that subglacial outbursts are a major component of the ice dynamical system. Including subglacial discharge of lakes into large-scale numerical ice sheet models

for Antarctic ice sheet evolution (paleo as well as future predictions) will become necessary.

The relation between ice dynamics, subglacial lake dynamics, and circulation is complicated and influences the mass balance at the interface between both systems. A first step in fully coupling both subglacial lake circulation and ice flow has been realized [*Thoma et al.*, 2010]. However, the processes involved are similar to those that govern any interaction between the ice sheet and a water body, such as grounding line migration and marine ice sheet stability. Therefore, subglacial lake discharge research will be crucial in future developments and improvements of cryospheric dynamics.

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