A conceptual model of supraglacial lake formation on debris-covered glaciers based on GPR facies analysis.

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| Complete List of Authors: | Mertes, Jordan; Michigan Technological University, Geological and Mining Engineering and Sciences; University Centre in Svalbard, Department of Arctic Geology
Thompson, Sarah; University Centre in Svalbard, Department of Arctic Geology
Booth, Adam; University of Leeds, School of Earth and Environment
Gulley, Jason; University of South Florida, School of Geosciences
Benn, Douglas; University of Saint Andrews, Department of Geography and Sustainable Development |
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Authors

Jordan R. Mertes\textsuperscript{1,2}, Sarah S. Thompson\textsuperscript{2}, Adam D. Booth\textsuperscript{3}, Jason D. Gulley\textsuperscript{4},
Douglas I. Benn\textsuperscript{5}

Title

A conceptual model of supraglacial lake formation on debris-covered glaciers based on GPR facies analysis.

Affiliations

\textsuperscript{1} Department of Geological and Mining Engineering and Sciences, Michigan Technological University, 630 Dow Environmental Sciences, 1400 Townsend Drive, Houghton, MI 49931, USA

E-mail: JRMertes@mtu.edu

\textsuperscript{2} Department of Arctic Geology, University Centre in Svalbard, Postboks 156, NO-9171, Longyearbyen, Svalbard, Norway

E-mail: Jordan.Robert.Mertes@unis.no

\textsuperscript{3} School of Earth and Environment, University of Leeds, Leeds, LS2 9JT, UK

\textsuperscript{4} School of Geosciences, University of South Florida, 4202 E. Fowler Avenue, NES 107, Tampa, FL 33620-5550 USA
Abstract

Supraglacial lakes and ponds can create hotspots of mass loss on debris-covered glaciers. While much research has been directed at understanding lateral lake expansion, little is known about the rates or processes governing lake deepening. To a large degree, this knowledge gap persists due to sparse observations of lake beds. Here we report on the novel use of ground penetrating radar (GPR) surveys to simultaneously collect supraglacial lake bathymetry and bottom composition data from Spillway Lake (surface area of $2.4 \times 10^5$ m$^2$; volume of $9.5 \times 10^4$ m$^3$), which is located in the terminus region of the Ngozumpa Glacier in the Khumbu region of the Nepal Himalaya. We identified two GPR bottom signals corresponding to two sedimentary facies of (1) sub-horizontal layered fine sediment drape and (2) coarse blocky diamict. We provide an understanding of the changes in subaqueous debris distribution that occur through stages of lake expansion by combining the GPR results with in situ observations of shoreline deposits matching the interpreted facies. From this, we present an updated conceptual model of supraglacial lake evolution, with the addition of data on the evolving debris environment, showing how dominant depositional processes can change as lakes evolve from perched lakes to multi-basin base-level lakes and finally onto large moraine-dammed lakes. Throughout lake evolution, processes such as shoreline steepening, lakebed collapse into voids and conduit interception, subaerial
and subaqueous calving and rapid areal expansion alter the spatial distribution and
makeup of lakebed debris and sediments forcing a number of positive and negative
feedbacks on lake expansion.

Introduction

Thick layers of debris on debris-covered glacier tongues have been found to insulate
debris-covered glaciers from mass loss in response to climate warming by decreasing
heat flux to the underlying ice surface (Østrem, 1959, Nakawo and Rana, 1999,
Nicholson and Benn, 2006, Reid and Brock, 2010, Reznichenko, et al., 2010, Shea and
Immerzeel, 2016). However, mass loss on some debris-covered glaciers has been
shown to be keeping pace with adjacent clean ice glaciers (Gardelle, et al., 2012, Kääb,
et al., 2012). This phenomenon may be attributed to irregular surface evolution due to
uneven sub-debris melting leading to the presence of features such as supraglacial
lakes and ice cliffs, which act as localized hotspots of melting. In some cases, these
hotspots have been shown to contribute to mass loss significantly, though only making
up a very small part of the glacier surface area (Sakai, et al., 2000b, Sakai, et al., 2002,
Through time however, the growth of supraglacial lakes and the rapid backwasting of
bordering ice cliffs, can eventually lead to coalescence with nearby lakes, and increased
chances to form potentially dangerous moraine-dammed lakes (Sakai, et al., 2000a,
Watanabe, et al., 2009).

Supraglacial lakes contribute to glacier mass loss by driving melt in the tongues of
stagnant, debris-covered glaciers where surface slopes are generally < 2° (Reynolds,

Much of what is known about the evolution of supraglacial lakes concerns areal expansion and the rates and fundamental processes underpinning areal lake expansion are well established from remote sensing and surface surveys (Mool, 1995, Quincey, et al., 2007, Komori, 2008, Fujita, et al., 2009, Watanabe, et al., 2009, Bajracharya and Mool, 2010, Gardelle, et al., 2011, Sawagaki, et al., 2012, Thompson, et al., 2012, Liu, et al., 2015, Wang, et al., 2015, Thompson, et al., 2016). Far less, however, is known about rates of, and processes controlling, lake deepening. To a large degree, this is because the remote sensing techniques used to measure lake depth on clean ice glaciers (Sneed and Hamilton, 2007, Fitzpatrick, et al., 2014), where lakes are clear, do not work in the turbid lakes that are characteristic of debris-covered glaciers. As a result, much of what is known about lake bathymetry has been determined from labor intensive field campaigns which have typically been conducted at irregular intervals.
While many details of lake deepening remain unknown, two physical processes are thought to dominate: subaqueous calving (Röhl, 2008) and direct lakebed melting (Chikita, et al., 1999, Sakai, et al., 2000b, Chikita, et al., 2001). Subaqueous calving has been observed to remove large volumes of ice from lake beds at a number of glaciers in New Zealand (Kirkbride and Warren, 1997, Warren and Kirkbride, 2003, Röhl, 2008). However, the mechanism relies in part on the buoyancy of ice and doubt remains as to the impact of the subaqueous debris layer (Benn, et al., 2012). Melt rates of lakebeds are critically dependent on water temperature as well as the presence or absence of lake bed debris. For example, on the Lirung Glacier, Nepal, low subaqueous melt rates of $< 0.01 \text{ cm d}^{-1}$ occurred in lake beds covered in sediments, but where lakes were floored with bare ice, subaqueous melt rates were $2-4 \text{ cm d}^{-1}$ (Miles, et al., 2016).

Improved understanding of the rates and processes of lake deepening requires a detailed temporal and spatial record of lake bathymetric evolution, as well as an understanding of debris distribution and physical characteristics. Most bathymetric survey techniques rely on weighted lines or hand-held electronic sonar devices (Yamada and Sharma, 1993, Kadota, 1994, Benn, et al., 2001, Sakai, et al., 2003, Yamada, et al., 2004, Sakai, et al., 2005, Fujita, et al., 2009, Sawagaki, et al., 2012, Thompson, et al., 2012, Somos-Valenzuela, et al., 2014, Horodyskyj, 2015, Thompson, et al., 2016). These techniques are time consuming and labor intensive, lack fine spatial resolution and provide little to no information about bottom compositions. Improved spatial resolution was obtained from surveys conducted using remotely operated underwater vehicles (ROVs) outfitted with sonar transducers (Horodyskyj, 2015), and sonar data can be analyzed to provide some information about bottom hardness and
texture (Kenny, et al., 2003) but most commercial sonar systems cannot provide
information about debris thickness.

Ground penetrating radar (GPR) is an underutilized tool for characterizing supraglacial
lake bathymetry as well as bottom composition and thickness. While GPR has been
used to measure glacier depths and ice volume for bathymetric and geological mapping
of non-supraglacial lakes (Haeni, et al., 1987, Sellmann, et al., 1992, Moorman and
2011, Sambuelli and Bava, 2012), few studies have used GPR to investigate lakes on
debris-covered glaciers (Reynolds, 2011).

In this paper we present the results of GPR surveys of Spillway Lake, Ngozumpa
Glacier, Nepal to demonstrate the potential of GPR as an investigative tool for
characterization of supraglacial lake bathymetry and lake bed characteristics. We then
use a combination of GPR facies analysis (e.g., Bristow, 1995, Ruffell, et al., 2004) and
field observations to characterize and interpret the subaqueous debris distribution.
Finally, we present an updated conceptual model of debris-covered glacier supraglacial
lake evolution that is based on the facies analysis and a review of published literature
on supraglacial lakes.

Study Site

Ngozumpa Glacier (27.97°N; 86.69°E) is located in the Khumbu Himal, Nepal, ~25 km
west of Sagarmatha (Mt. Everest). The Ngozumpa Glacier flows down from Gyachung
Kang (to the northeast) and Cho Oyu (to the northwest) (Fig 1). At ~20 km long, the
glacier is one of the largest in Nepal, and the lower 15 km is extensively debris covered.
Debris thickness is highly variable but generally increases from a thin, discontinuous cover just below the equilibrium line altitude (~5700 m) to the terminus, where it is continuous (excluding ice cliffs) and 1-3 m thick (Benn, et al., 2012, Nicholson and Benn, 2012). This debris layer has affected the glacier mass balance and altered the distribution of ablation relative to a clean ice glacier; the maximum ablation now occurs up glacier from the terminus where the debris layer is thin, rather than at the terminus. This ablation pattern has caused a reduction in surface gradient and an associated decrease in the driving stresses (Bolch, et al., 2008, Quincey, et al., 2009). In recent decades the glacier has experienced substantial surface lowering, and, in the ablation area, the surface now lies >150 m below the Little Ice Age moraines (Sakai and Fujita, 2010). Decreases in surface gradient and driving stresses have ultimately led to a reduction in glacier velocity which has caused the lower ~7 km of the glacier to become stagnant (Quincey, et al., 2009, Thompson, et al., 2016).

The glacier has highly variable surface topography, consisting of large hummocks, ridgelines, lakes and ice cliffs. The surface relief is approximately 10s of meters around larger hummocks and lake ice cliff contacts. The mean surface slope of the lowest 5km has been measured previously at ~2° (Hands, 2004). This slope is within the range cited as having potential for the formation of extensive supraglacial lakes (i.e. ≤2°) (Reynolds, 2000, Quincey, et al., 2007, Röhl, 2008) and in the lowest ~2km where our study focuses, the overall slope decreases to <1°.

In the early 1990s a collection of lakes and ponds began to form at the hydrological base level of the Ngozumpa Glacier (Thompson, et al., 2012). Since this time, the lake
basins have continued to grow and coalesce to cover an area of $2.4 \times 10^5$ m$^2$ with a
volume of $9.5 \times 10^4$ m$^3$ in 2014 (Thompson, et al., 2016). This growing lake body is,
informally, referred to as Spillway Lake. The site has been the focus of a series of
studies using remote sensing, sonar measurements and extensive field mapping
Horodyskyj, 2015, Thompson, et al., 2016). Our new surveys, presented here, focus on
two connected basins within the Spillway Lake system, adjacent to the eastern lateral
moraine and the central lake basin (Fig 2 A, B).

Methods

During December 2014, five GPR transects were surveyed on the frozen surface of
Spillway Lake, using a Malå ProEx Professional Explorer GPR system linked to an
unshielded Rough Terrain Antenna (RTA) of 25 MHz center frequency. The RTA is
roughly 13 m long and houses a single pair of transmitter/receiver antennas separated
by 6 m in an inline configuration. During acquisition, the system was connected to a
Leica Differential Global Positioning System (DGPS) rover, and set to record a trace
every 0.3 s. The ProEX unit was placed inside an inflatable kayak behind which the
RTA was towed. The system was towed on foot at an approximately constant speed of
~1 m s$^{-1}$. The interval between recorded traces was $0.26 \pm 0.06$ m, regularized (in
Mathworks Matlab® software) to 0.2 m during processing. A bulk correction of 11 m
(applied in the opposite direction to the profile heading) was applied to all positions to
remove the offset between the DGPS antenna and the midpoint between the GPR
antennas. Due to a battery failure in the Leica base station, GPS measurements cannot
be post processed and corrected to centimeter accuracy. Therefore, to assure the most
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accurate transect positioning, we have used field video and photographic data to verify correct positioning of transect start-stop positions relative to high-resolution satellite imagery (~ 0.48 m pan-sharpened GeoEye-1) where possible. Uncorrected DGPS data from sonar surveys, conducted prior to the GPR survey, show errors on the scale of ±1-2 m in the x-y plane.

The effective depth of investigation of a GPR signal is substantially impeded in electrically conductive ground (e.g., high clay fraction and/or saline water content) as the electromagnetic energy of the GPR wavelet is more rapidly absorbed (Annan, 2004). Groundwater is therefore often considered an impediment to GPR surveying.

Supraglacial lakes, however, can have very low electrical conductivities, with conductivities of ~2.3 mS m\(^{-1}\) to ~12 mS m\(^{-1}\) being reported from supraglacial lakes on debris-covered glaciers (Watanabe, et al., 1995, Sambuelli and Bava, 2012). These conductivities are close to pure water and lakes should thus be transparent to radar.

The RTA used to conduct our GPR surveys allows for rapid data acquisition, but it does not allow common midpoint (CMP) surveys to be conducted for velocity analysis. As a result, GPR surveys reported here do not provide information about lakebed debris thickness. They can, however, provide information about bottom composition and water depth based on the first arrival of the GPR signal and its character immediately thereafter. In the absence of CMP-derived velocities, we assume a relative dielectric permittivity, \(\varepsilon_r\), for water of 80 (Annan, 1992). Surface ice cover is ~0.2 m thick; therefore, our time-to-depth conversions assume a GPR velocity of 3.3 \times 10^{-2} \text{ m s}^{-1} throughout. While \(\varepsilon_r\) can vary with temperature (e.g. Malmberg and Maryott (1956)) measured \(\varepsilon_r \approx 88\) at 0°C, the uncertainty in depth estimates is expected to be small and
the difference in derived velocity, and therefore depth estimates, between $\varepsilon_r$ values of 80 and 88 is < 5%. Hence, in the absence of lake water temperature measurements, we use $\varepsilon_r = 80$. We estimate the wavelength of the 25 MHz RTA wavelet to be 1.3 m; the $\frac{1}{4}$-wavelength criterion for vertical and (following migration) lateral resolution suggests that our limit of resolution is 0.3 m. Data processing steps, implemented in Sandmeier ReflexW® 7.5 software, are listed in Table 1. The data generally respond well to the imposed processing flow, although artefacts of normal moveout correction remain at early travel-times where the propagation path of GPR energy is comparable in length to the 6 m offset between the two antennas. Very few artefacts of migration remain; those that do are presumably caused by off-line scatter, and cannot be correctly migrated given the 2D acquisition.

By picking the first-break of lakebed reflections, we calculated bathymetry for each trace of the GPR dataset. This operation was straightforward in the deepest sections of the lake but was subject to greater uncertainty in shallower sections where the bed reflection was generally more complex. Bed reflection characteristics are discussed in the Results section under “GPR Facies Classification”. In addition, the shallower areas are subject to distortions in our normal moveout correction (e.g. Fig 3, beginning of transect 1, A). Given the finite lateral resolution of the GPR wavelet, we picked bathymetry to be deliberately smooth, and estimate that the uncertainty in the calculation of lake depth to be at maximum ±1 m.

The performance of the GPR depth conversion is assessed against the results of a separate sonar bathymetry point survey performed in December 2014 using a Hummingbird 385ci DualBeam™ sonar (Thompson, et al., 2016). The grid spacing of
sonar survey was ~20 m along track and 6-40 m between tracks (Fig 2), and results
were spline interpolated with a shoreline polygon barrier in ESRI ArcMap to provide a
map of lake bathymetry. However, because the sonar bathymetry survey was not
conducted specifically for comparison with the GPR survey, the GPR survey lines do
not directly overlap the sonar points. Thus, we expect there to be variability in the
comparison and use this as a broad assessment of the GPR depth conversion, rather
than an assessment of the superiority of methods. Sonar points located closer to the
GPR profiles are likely to provide a better corroboration of the GPR data than those
located at a greater distance. The source pulse of the echo-sounder blends frequencies
of 85 and 200 kHz, which gives vertical and lateral resolution comparable to the GPR,
assuming a sonic velocity of 1500 m s\(^{-1}\) in water. We therefore extracted interpolated
sonar depth estimates along the radar transects to provide the necessary comparison
between the two platforms.

Finally, we used GPR facies analysis to classify the lakebed from the processed GPR
data. Facies analysis (Bristow, 1995, Ruffell, et al., 2004) is a qualitative means of
characterizing styles of reflectivity, thereby allowing the mapping of consistent features,
which have similar reflection geometry. We applied this qualitative approach to
classification, as a more quantitative analysis is not possible in the absence of CMP
data.

Results

GPR bathymetry
In GPR transects from basin A (Fig 3; transects 1-4) and basin B (Fig 4), we interpret
the first arrival after a transparent section of data to be the reflection from the lakebed.
The maximum two-way travel-time of the lakebed reflection is ~ 1300 ns, corresponding
to a maximum lake depth of ~ 21 m (Fig 3, transect 1). In Basin A, the lakebed
geometry appears generally bowl-shaped, relatively flat in the deeper areas, sloping up
to the shoreline at an angle of 20-30° relative to the lake surface. A prominent hummock
occurs in the center of the basin in transects 1 and 3 (Fig 3). The hummock is inferred
to be ~60 m wide at its base, with a topographic high point at a depth of ~ 9 m, ~ 16 m
shallower than the surrounding lakebed. Transect 2 (Fig 3) also shows evidence of the
hummock, but the shoreward slope is not well defined. In Basin B, lakebed geometry
shows a gradual deepening from 1 m in the south to 6 m in the north over a distance of
~120 m, followed by rapid deepening to 15 m (Fig 4). The lake shallows again to depths
of 3-7 m between distance of 220 m and 260 m along track. This section of the lake
appears to have a more undulating bathymetry.

Comparison with sonar data

Only six of the sonar survey points are located within a distance of ≤1 m from the GPR
transects. The mean difference between these sonar point depth measurements and
the closest measured GPR depths was 1.4m (Fig 5a). Differences between the GPR
depths and the interpolated sonar depths were generally larger, with positive differences
ranging from 2.3 to 7.7 m and negative differences ranging from -1.1 to -6.0 m (Fig 5b).
Transects 2A, 4A and 1B, which roughly follow grid lines of the sonar measurements, all
have differences less than ±5m with mean differences and standard deviation of 1.2±1.2
m, -0.7±1.7 m and -0.8±1.4 m respectively (Fig 5b). Conversely, transects 1A and 3A,
running oblique to sonar measurement grid, all have differences greater than 5 m and
less than -4.3 m with means and standard deviations of -0.4±3.0 m and 1.2±3.8 m
respectively (Fig 5b). The mean difference of all transects is 0.1 m, with a standard
deviation of ±2.5 m.

GPR facies classification

Sediment appears to cover the entire surveyed parts of the lakebeds. Two facies can be
distinguished based on the characteristics of the lakebed reflectors (Fig 6). Facies 1
shows coherent, sub horizontal reflectivity, with parallel events either side of incoherent
responses (Fig 6a). Facies 1 is present in the deeper areas of both lake basins and
around the lake flanks (Fig 3 and 4). Facies 2 shows low-amplitude, chaotic reflectivity,
characteristic of structural complexity (Fig 6b). Facies 2 is present mostly around the
submerged hummocks in basin A as well as the eastern shallows of transect 4 A (Fig
3).

Interpretation and discussion

GPR analysis and interpretation

Differences in point depth measurements made using GPR and sonar are smallest
where distances between the measurements were smallest (Fig 5a). Similarity of depth
measurements results from the GPR and sonar having similar un-migrated horizontal
resolutions, with a mean difference of 0.52 m over the depth ranges in our study.
Differences between GPR depths and interpolated sonar depths are likewise smallest
along transects that have the largest number of proximate sonar point measurements,
as well as where lake bathymetry is relatively flat. For example, transects 2A, 4A and
1B appear very similar to the interpolated sonar depths (Fig 7). All three transects have
a gradually changing bathymetry, with only a few regions of rapid change. GPR traces
for these transects are located a mean distance of 8.4±2.8 m, 5.3±1.7 m and 7.2±3.5 m
away from their nearest sonar point measurements respectively (Fig 8). The maximum
and minimum distances for transects 2A and 4A are 12.9 – 2.6 m and 8.9 – 1.6 m
respectively. Transect 1B, however, has a much larger range of distances with a
maximum and minimum of 17.1 – 0.9 m. The mean distance to sonar measurements
along Transects 1A and 3A is similar to that for Transect 2A, but the variability along
these two transects is greater than for 2A, with maximum distances at 20.4 m and 20 m
respectively (Fig 8).

The two sets of depth measurements become more divergent where lake bathymetry is
more complex (e.g. Figure 7, transect 3A, 50-150 m) likely as the result of sparse point
data to constrain interpolation. For example, the greatest differences between depth
measurements occur in transects 1A and 3A, where point data are spaced further than
13 m apart (e.g. Fig 2 red ellipse) and interpolation errors are likely to be the greatest.

Similarly, transect 1B shows small variation in differences, yet has numerous outliers
(Fig 5b) that correspond to large distances from sonar point measurements (Fig 8).

These locations of larger difference are easily identifiable in Figure 7 and seen as the
central area of basin B where data are again sparse. Small differences in depth
measurements using the two techniques indicate GPR provides depth information that
is roughly equivalent to sonar, but GPR has the added advantages of rapid data
acquisition when collected on frozen lake surfaces and provides information about lakebed composition and structure.

**Facies**

Facies 1 is consistent with a more uniform scattering in which discontinuities are much smaller than the radar wavelength (i.e. 1.3 m). We interpret these sub-parallel sequences of reflectivity as layers of fine sediment, similar to the bedded sand deposits exposed around some lake margins (Fig. 9, 1&2 (right)). By contrast, Facies 2 is consistent with a chaotically scattering lakebed where discontinuities are on a spatial scale comparable with the radar wavelength (i.e., 1.3 m). We interpret Facies 2 as areas of the lakebed covered with coarse debris, including boulders and large clasts, similar to the widespread blocky diamict debris cover exposed on the glacier surface (Fig. 9, 1&4). Our facies classifications and identification agree well with the mapped distribution of bottom surface types found by Horodyskyj (2015), and are also consistent with the observation that bathymetry associated with Facies 1 is likely to have less abrupt changes than in areas where Facies 2 is found.

The lack of exposed ice in our lakebed surveys suggests that any bare ice in lake basins is rapidly covered by sediments. Facies sequences, however, suggest that bare ice is exposed at lakebeds at least transiently. For example, if Facies 2 is in contact with ice on the lakebed, it cannot be known if the sediments were deposited as part of a mass wasting event onto a bare ice bed following lake formation or if they reflect an original debris-covered ice surface that was submerged following surface downwasting. In contrast, because Facies 1 suggests deposition by settling of finer grained materials,
the superposition of Facies 1 with glacier ice at the lake bed strongly suggests the lake bed consisted of bare ice likely present prior to Facies 1 deposition (Figs 9, 3&4; Fig 10). Facies relationships suggest that up 20% of our survey lines were initially composed of bare ice, but were later covered by Facies 1 sediments.

Overlapping facies relationships demonstrate that the dominant sediment/debris delivery mechanisms to lakes can switch from mass wasting, toppling from ice cliffs and/or sliding from moraine ridges to settling of suspended fines, creating a transition from Facies 2 to Facies 1. For example, the upward transition from Facies 2 to Facies 1 (Fig 3) between 130 and 160 m in transect 3A, and between 120 to 160 m in transect 2A, indicate that the dominant sediment delivery mechanism switched from mass wasting to deposition of fines. Such a transition would be expected to occur as lakes expanded. Sediments would initially be deposited via mass wasting occurring as ice cliffs retreated around the lakeshore, causing small landslides of sediments into lakes. As lake area expanded, and sources of sliding debris move further away, sedimentation changes to settling of fines.

Transitions from Facies 1 to Facies 2 require a different explanation. We did not find distinct evidence for this transition in our GPR data, but it was observed in sediment packages associated with a drained lake basin ~60 m north of lake basin A (Fig 2, white hexagon 3; Fig 9, 3). Based on freely available declassified KH-9 and Landsat imagery of the AOI, it can be seen that between 1964 and 1967 a perched lake and backwasting ice cliff located ~60-70 m north in 1964 was expanding and most likely intercepted this location. Throughout the mid-late 1970s the area appears to be submerged but towards the beginning of the early 1980s it is once again exposed. As Spillway Lake began to
form and expand, this location was once again submerged, sometime in 2005-2006, and subsequently was re-exposed in the 10/15/2009 Landsat image. A stratigraphic sequence of fine-grained sediments deposited near-parallel with a bare ice surface was exposed over an ice cliff at the lakeshore, indicating deposition onto submerged bare glacier ice. There is an abrupt change from fine sediment to a coarse-grained diamict wedge 1.5 m thick, which extends part way across the underlying fines. Fine-grained material has also been deposited on top of the diamict, supporting the characterization of this sequence in the lake environment, and was the result of a subaqueous debris slumping event (Fig 10).

A conceptual model of supraglacial lake evolution for debris-covered glaciers

Here we present a conceptual model of supraglacial lake evolution, the basis of which is a consolidation of the literature investigating lake formation and expansion, with the addition of the facies relationships observed in our study. Facies transitions suggest that the spatial distribution of debris and depositional processes change as supraglacial lakes grow. These changes reflect varying environmental factors such as proximity to ice cliffs and debris-covered slopes, initiation of calving, interception with englacial conduits or voids, and subaqueous and subaerial mass movement events. These relationships provide the addition of the evolving debris environment to the model of an expanding supraglacial lake, highlighting the changing debris-related constraints and feedbacks as a supraglacial lake evolves.

During early stages of lake growth, water accumulates in small, isolated lake basins that are perched above base level and not connected to supraglacial streams or englacial
conduits (Benn, et al., 2001) (Fig 11, A). The surrounding shoreline likely consists primarily of low gradient, stable debris-covered slopes. Prior to infilling, the lakebed is likely composed of similar diamict as is found above the lake surface (Facies 2). During lake infilling, sedimentation rates in the lake would be low, because sediments would be sourced primarily from windblown dusts or fine-grained materials leached from the surrounding diamict by percolating meltwaters (Facies 1). Consequently, Facies 1 is deposited over Facies 2.

Expansion of small, perched lake basins occurs by slow melting of subaqueous ice around the lake margins, or possibly collapse into englacial voids, which steepens the lakebed slopes, eventually forming steep bare ice faces or ice cliffs (Benn, et al., 2001, Röhl, 2008, Thompson, et al., 2012) (Fig 11, B). Bare ice faces and ice cliffs increase rates of lake expansion and trigger a change in deposition. Bare ice faces often have a thin coating of dark, wet mud which reduces the surface albedo and increases absorption of solar radiation, creating localized hot spots of enhanced melting (Benn, et al., 2001). Ice cliffs can lead to rapid lake expansion through calving, as thermo-erosional notching at the waterline leads to fracturing (Benn, et al., 2001, Röhl, 2008, Miles, et al., 2016). Lake expansion by melt retreat of steep, bare ice faces and calving would cause Facies 2 type sediments to be deposited by slumping around the expanding lake margin, with deposition of fine-grained Facies 1 sediments in the central deeper areas of the lake.

Sediment deposition can force a positive feedback on lake expansion. Slumping provides a constant influx of diamict, which helps keep turbidity high. This increase in turbidity can decrease lake surface albedo, thereby increasing absorbed solar radiation,
which keeps surface temperatures high. Conversely, a negative feedback results from deposited fine sediments (Facies 1) on the lakebed insulating underlying ice from melt, decreasing bottom water cooling. These effects combine to promote melting along lake perimeters, thereby enhancing areal expansion.

Perched lake expansion and sedimentation is halted when lakes intercept englacial conduits or sediment filled crevasse traces and rapidly drain (Fig 11, C) (Benn, et al., 2001, Wessels, et al., 2002, Gulley and Benn, 2007, Röhl, 2008, Benn, et al., 2009, Benn, et al., 2012, Thompson, et al., 2012). While conduits temporarily drain small perched lakes, they play an important role in glacier surface lowering and the development of larger lakes (Kirkbride, 1993, Immerzeel, et al., 2014, Thompson, et al., 2016). Conduit collapse creates new depressions with new bare ice faces where enhanced melting can continue and lakes may eventually be recharged with water from up glacier. In addition, water transported from a lake drainage event loses heat to conduit walls, thereby adding to mass loss through internal ablation (Kirkbride, 1993, Benn, et al., 2012, Miles, et al., 2016, Thompson, et al., 2016) and promotes future collapse by widening conduits (Gulley and Benn, 2007). The presence of Facies 1 in direct contact with glacier ice may be due to a collapse and drainage event that flushed lakebed sediments out of the lake basin, leaving a bare ice surface exposed to melt and sedimentation.

Cycles of lake expansion, drainage by conduits and collapse, and refilling allow smaller, isolated lakes to coalesce behind the terminal moraine to form a large, moraine-dammed lake; examples include Imja Tsho, or Tsho Rolpa (Nepal Himalaya) (Fig 11, D-F). While the physical processes driving the change from a series of perched lakes, like
Spillway Lake reported here, to a large moraine dammed lake remain poorly understood, it is thought that coalescence of small lakes into a larger one triggers rapid lake expansion through subaerial slab calving, and increased subaqueous waterline melting (Kirkbride, 1993, Sakai, et al., 2009) (Fig. 11, E). While slab calving of exposed ice cliffs can cause rapid lake expansion, deposition of subaerial debris located along the top of ice cliffs onto the lakebed can force a negative feedback on lake expansion (Fig. 11, E). Accumulation of debris at the base of ice cliffs has been observed to separate the ice cliff from the lake, halting calving and lake expansion at that location (Thompson, et al., 2016).

Lake-bed disintegration has been proposed as a key trigger to initiation of full front calving and rapid retreat (Kirkbride, 1993). However, more recent work suggests rapid lake expansion by calving does not require the lake to deepen to the glacier bed (Fig. 11, F). Robertson et al. (2012) observed subaqueous debris layers between 5-10 m in thickness at the calving margins of Mueller, Hooker and Tasman glaciers. Using a CHIRP sonar, they were able to visualize the calving margin where they found a debris-covered sloping ice foot projecting into the lake at an angle of about 40º, indicating the calving front did not reach to the glacier bed (Robertson, et al., 2012). Similarly, a GPR study at Imja Tsho, found glacier ice just up-glacier from the calving front was ~80 m deeper than the deepest point in the lake, indicating that glacier ice was present below the lake bed (Somos-Valenzuela, et al., 2014). Despite debris cover, data from Imja Tsho indicates that some lakebed deepening still occurs. Taking into consideration that the lake level has been stable for the past decade, at roughly 5010 m a.s.l., the mean rate of deepening at Imja Tsho (2002-2012) was roughly 0.86 m/yr, with maximum
depths increasing from 98 m to 116 m (Somos-Valenzuela, et al., 2014). Deepening was faster near the calving front and slower nearer the terminus. Some of this deepening may be due to sub-debris melting, but some is due to subaqueous calving (Somos-Valenzuela, et al., 2013, Somos-Valenzuela, et al., 2014).

In summary, using the GPR facies relationships identified in our surveys with field observations, we were able to infer feedbacks between processes controlling supraglacial lake growth and sediment deposition. We have developed a conceptual model of how dominant depositional processes can change as lakes evolve from perched lakes to multi-basin base-level lakes and finally onto large moraine-dammed lakes. Throughout lake evolution, processes such as shoreline steepening, lakebed collapse into voids and conduit interception, subaerial and subaqueous calving and rapid areal expansion alter the spatial distribution and makeup of lakebed debris and sediments which, in turn, can control rates of deepening by enhancing or diminishing heat conduction to the underlying ice.

Conclusion

The results of our GPR surveys not only provide high resolution bathymetric information (beneath the survey lines) allowing us to map the morphology of the lake bed but the additional facies interpretation provided a detailed lakebed debris characterization. This work demonstrates the applicability of GPR as a tool for supraglacial lake investigation and monitoring. We found that lake depth surveys can be completed rapidly with GPR when lake surfaces are frozen. While studies of supraglacial lakes have traditionally relied primarily on sonar, we showed that GPR provides depth information that is
equivalent to sonar, with added information about sediment types on lakebeds. Facies analysis highlights the additional information that can be derived from lakebed surveys using GPR. Sonar data can provide some information about bottom composition, such as surface hardness and texture (Horodyskyj, 2015) but cannot provide the facies information as reported here. Selecting GPR units with CMP survey capability would further expand the utility of GPR, as CMP surveys can provide measurement of the velocity through the debris, allowing the calculation of sediment layer thicknesses, an important parameter for modelling subaqueous heat flux and lake deepening. Further, conducting paired CMP surveys would allow for debris thicknesses to be mapped by determining the material velocity of different facies types, ultimately leading to a much more complete understanding of the spatial distribution, quantity and makeup of lakebed debris.

Future investigations should aim to perform repeat 3-D GPR surveys, which will allow not only detailed changes in lake bathymetry and bed morphology to be measured but also changes in debris distribution. Knowledge such as this is of paramount importance for modelling potential subaqueous lake expansion and understanding the specific thresholds that trigger rapid growth from supraglacial lakes to moraine-dammed lake.

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Tables

Table 1. Processing operations and parameters applied sequentially to profiles of GPR data.

<table>
<thead>
<tr>
<th>Operation</th>
<th>Parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dewow filter</td>
<td>Window length = 80 ns</td>
</tr>
<tr>
<td>(removal of low-frequency component of trace)</td>
<td></td>
</tr>
<tr>
<td>Static correction</td>
<td>All first-breaks synchronized to 20.6 ns (= 0.3 m/ns x 6.2 m)</td>
</tr>
<tr>
<td>(alignment of time-zero in traces)</td>
<td></td>
</tr>
<tr>
<td>Regularization of trace interval</td>
<td>Trace interpolation to 0.2 m</td>
</tr>
<tr>
<td>Normal moveout correction</td>
<td>Assumed velocity = 0.033 m/ns</td>
</tr>
<tr>
<td>Ormsby bandpass filter</td>
<td>Corner frequencies = 4-8-50-100 MHz</td>
</tr>
<tr>
<td>Constant velocity Kirchhoff migration</td>
<td>Migration aperture = 40 m</td>
</tr>
<tr>
<td></td>
<td>Velocity = 0.033 m/ns</td>
</tr>
<tr>
<td>Median filter</td>
<td>Window = 3 traces x 3 samples</td>
</tr>
<tr>
<td>Amplitude gain</td>
<td>Gain function based on constant velocity</td>
</tr>
<tr>
<td></td>
<td>geometrical spreading correction</td>
</tr>
<tr>
<td>------------------</td>
<td>----------------------------------</td>
</tr>
<tr>
<td>Depth conversion</td>
<td>Constant velocity = 0.033 m/ns</td>
</tr>
</tbody>
</table>
Figure 1: Overview Geoeye satellite image of Ngozumpa Glacier with Spillway Lake complex indicated in black rectangle.

279x512mm (300 x 300 DPI)
Figure 2: Spillway Lake (basins A & B) area of interest shown with interpolated depth map and overlay of sonar point depth locations and GPR transects. Red ellipse indicates area of sparse sonar point measurements along transect 3, A. White hexagons around numbers indicate locations of sediment samples seen near shoreline in Figures 9 & 10.

190x168mm (300 x 300 DPI)
Figure 3: GPR transects 1-4, (basin A), post processed with dashed black line indicating lakebed and labeled multiples and locations of inferred facies. Vertical dotted line indicates crossing point of transects 1 and 3. Vertical exaggeration is roughly 5x. Direction of travel is from left to right.
Figure 4: GPR Transect 1, (basin B), post processed with dashed black line indicating lakebed and labeled multiples and locations of inferred facies. Vertical exaggeration is roughly 5x. Direction of travel is from left to right.

101x52mm (300 x 300 DPI)
Figure 5: GPR depths compared to sonar point depths and interpolated sonar depth (A). Boxplot of differences between GPR depths and interpolated sonar depth (B). Transects 2A, 4A and 1B display a much smaller interquartile range indicating better agreement between depths, with averages from 1.3 - 1.4m (x).
Figure 6: Subsets of distinct radar facies. (A) Coherent, sub-horizontal reflectivity, often comprising subparallel sets of reflections. (B) Typical low-amplitude and chaotic reflectivity, prone to migration noise suggesting structural complexity. Both facies plotted with equivalent amplitude scales.

83x52mm (300 x 300 DPI)
Figure 7: Comparison of GPR depths (gray dashed) and interpolated sonar depths (solid black), with differences (top dash-dot-dot). Note the areas of large difference in transects 1A and 3A, corresponding to areas where there is little overlap between the methods.
Figure 8: Boxplot of distance from GPR trace to nearest sonar point measurement.

166x133mm (300 x 300 DPI)
Figure 9: Examples of exposed shoreline facies marked in Figure 2 as white hexagons.

1. Mixed coarse blocky material with occasional evidence of sedimentation. Mainly Facies type 2 shoreline transitioning to Facies type 1 towards right side of picture.

2. Relict lake bed sediments showing strong Facies type 1 shoreline bordered by more blocky Facies 2 material.

3. Close up sediment sequence Facies type 1-2-1 from top to ice interface. Thick diamict wedge thins to right.

4. Thick diamict Facies type 2 overlying more type 1 sediment in contact with ice. Note thinner diamict layer towards bottom (red).
Figure 10: Overview of exposed large diamict wedge deposited between layered fines indicating debris redistribution onto relatively flat, sediment covered lakebed, followed by more fine deposition (Fig 2&9 example 3).

62x18mm (300 x 300 DPI)
Figure 11: A conceptual model of supraglacial lake evolution on debris covered glaciers with additional changes in sediment depositional processes affecting lakebed spatial debris distribution. (A) Isolated perched lakes, not connected to supraglacial streams or englacial conduits. (B) As perched lakes expand, debris slumping becomes more likely. (C) Some perched lakes may drain due to intersection with englacial conduits, potentially evacuating some lakebed sediment. (D) Ice cliff expansion and lake bed deepening lead to the intersection of the largest lake with a base level conduit. (E) Continued expansion of lakes in the area cause all to connect either through surface drainage networks or near surface conduits. (F) Lake expansion and coalescing leads to the formation of a single base-level moraine dammed lake.

153x79mm (300 x 300 DPI)