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Entrained Water in Basal Ice Suppresses Radar Bed-Echo Power at Active Subglacial Lakes

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Key Points:
- Active subglacial lakes, identified by surface deformation, do not create the expected bright and specular radar reflection
- Entrained water in basal ice suppresses radar power by scattering and attenuation, and it also likely alters the basal ice mechanics
- Understanding the radar expression of subglacial water on Earth provides context for investigations of subsurface water on planetary bodies

Supporting Information:

Supporting Information may be found in the online version of this article.

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Abstract Subglacial lakes have been mapped across Antarctica with two methods, radio-echo sounding (RES) and ice-surface deformation. At sites where both are coincident, these methods typically provide conflicting interpretations about the ice-bed interface. With a single exception, *active* subglacial lakes identified by surface deformation do not display the expected flat, bright, and specular bed reflection in RES data, characteristic of *non-active* lakes. This observational conundrum suggests that our understanding of Antarctic subglacial hydrology, especially beneath important fast-moving ice streams, remains incomplete. Here, we use an airborne RES campaign that surveyed a well-characterized group of active subglacial lakes on lower Mercer and Whillans ice streams, West Antarctica, to explore inconsistency between the two observational techniques. We test hypotheses of increased scattering and attenuation due to the presence of an active subglacial lake system that could suppress reflected bed-echo power for RES observations in these locations, finding that entrained water is most plausible.

Plain Language Summary The bottom of an ice sheet is insulated from cold air temperatures, often warm enough to melt and pond liquid water into lakes. These lakes beneath the ice sheet have been identified by two independent measurements, first with radar methods and second with changes in height of the ice surface (altimetry). Interestingly, the two methods rarely identify the same lakes: radar generally detects lakes in the ice-sheet interior, whereas altimetry detects *active* lakes near the ice-sheet margins that fill and drain within the time series of repeated measurements (~years). In this study, we investigate a group of active subglacial lakes at which both radar and altimetry data sets are available. We demonstrate that the radar returns from active lake reflections are much dimmer than expected based on non-active lake signatures and investigate the physical processes controlling those dim reflections. We argue that water moves into the ice when the lake fills or drains and that is the most plausible explanation for the observational discrepancy.

1. Introduction

The Antarctic ice sheet has a complex pattern of ice flow in which slow-moving interior ice feeds fast-moving ice streams and outlet glaciers (e.g., Mouginot et al., 2019; Rignot et al., 2011). These fast-moving ice streams account for as much as 90% of total ice discharge to the ocean (Bamber et al., 2000), and their high velocities are thought to be accommodated by the presence of meltwater at or near the bed (e.g., Engelhardt & Kamb, 1997; Tulaczyk et al., 2000). Understanding the relationship between subglacial water, basal lubrication, and ice dynamics is therefore critical to constraining Antarctica's mass balance and contribution to sea level change.

Until the mid-2000s, water beneath the Antarctic ice sheet was hypothesized either to be part of a steady-state transport system required to maintain fast flow (e.g., Engelhardt & Kamb, 1997; Joughin et al., 2004) or to reside in isolated subglacial lakes (e.g., Kapitsa et al., 1996). Subglacial lakes were identified by their flat, bright reflections in airborne and ground-based radio-echo sounding (RES) observations (e.g., Oswald & Robin, 1973). RES methods have been extremely effective for cataloging subglacial lakes throughout the continent, with over 250 discovered between 1973 and 2012 (Wright & Siegert, 2012). Some of those lakes have been inferred to be volumetrically stable for centuries to millennia (e.g., Bell et al., 2002).

More recently, a series of three independent studies identified clusters of spatially coherent surface-height displacements which were interpreted as the surface expression of subglacial water movement in and out of individual lake basins (Fricker et al., 2007; Gray et al., 2005; Wingham et al., 2006). Water was inferred to move rapidly (~months to years) through these lakes, with drainage that occurred as steady outflow or as an episodic

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release of water. These dynamic lakes with fluctuating water volumes have since been referred to as "active" subglacial lakes and provide a stark contrast to previously identified, more stable, "non-active" subglacial lakes. As a result of their clear surface expression, active lakes could be effectively mapped across Antarctica using satellite altimetry (e.g., B. E. Smith et al., 2009). Whereas non-active lakes were located primarily under slow, interior ice, these active lakes (of which there are over 100; Livingstone et al., 2022) tended to be located beneath fast-flowing ice streams and outlet glaciers (e.g., Fricker et al., 2016; Siegfried & Fricker, 2018). Linkages between filling and draining of nearby lakes indicated that a widespread, complex, dynamic hydrologic system likely exists beneath most of fast-moving regions of the Antarctic and Greenland ice sheets (Andersen et al., 2023; Fricker & Scambos, 2009; Neckel et al., 2021; Siegfried & Fricker, 2021; Wingham et al., 2006). These inter-connected systems of lakes can alter regional ice dynamics through water redistribution (Scambos et al., 2011; Siegfried et al., 2016; Stearns et al., 2008).

Multiple field campaigns have targeted active lakes for RES surveying but have repeatedly failed to produce the expected flat, bright signature of water at the ice-bed interface (e.g., Christianson et al., 2012; Siegert et al., 2014; Wright & Siegert, 2012; Wright et al., 2014). These studies have proposed a variety of hypotheses to account for the conflicting observations, none of which have yet been accepted, as we detail below. In some areas, even subglacial lakes which are assumed to be non-active have been difficult to interpret based on their radar echoes alone, leading to debates on both lake history and radar-echo signatures. South Pole Lake, East Antarctica, has been regarded as a "fossil lake" (Price et al., 2002), as freezing from a prior ice-dynamic regime (Beem et al., 2017), and as thermodynamically stable (Hills, Christianson, Hoffman, et al., 2022). Radar echoes beneath Devon Ice Cap were originally interpreted as hypsersaline lakes (Rutishauser et al., 2018, 2022), but additional investigations showed no evidence of subglacial water (Killingbeck et al., 2024). Radar echos from beneath Mars' southern ice cap have been interpreted as lakes (Lauro et al., 2021; Orosei et al., 2018), but are debated as possibly associated with some other conductive body rather than liquid water (Bierson et al., 2021; Lalich et al., 2022; I. B. Smith et al., 2021; Tulaczyk & Foley, 2020). From Earth to Mars and for upcoming planetary missions (Roberts et al., 2023) we must be able to reconcile these observations when their interpretations conflict in order to confidently quantify water and understand hydrological processes.

Here, we analyze an extensive airborne RES survey in a region of West Antarctica with inter-connected active subglacial lakes. These lakes are well characterized by altimetry data sets, and one is known to be full of water at the time of RES surveying through direct access (Tulaczyk et al., 2014). We interpret RES reflections at and near the ice-bed interface, using them to test five distinct hypotheses for suppression of RES bed-echo power over active subglacial lakes.

2. Methods

2.1. Study Site

The lower confluence of Mercer and Whillans ice streams, West Antarctica, informally referred to as Whillans Ice Plain, has an extensive and well-characterized system of eight inter-connected active subglacial lakes that have been analyzed via satellite observation (e.g., Fricker & Scambos, 2009; Fricker et al., 2007; Siegfried & Fricker, 2018) and ground-based geophysical suveying (e.g., Christianson et al., 2012; Horgan et al., 2012). Although the entire observational record of active subglacial lakes in Antarctica only spans 20 yrs (Siegfried & Fricker, 2018), active subglacial lakes on Whillans Ice Plain have undergone repeated fill-drain cycles on timescales ranging from months (Whillans Subglacial Lake; SLW) to 6 to 7 yrs (Conway Subglacial Lake). The only two Antarctic subglacial lakes to have been directly accessed and sampled—SLW (accessed 27 January 2013; Tulaczyk et al., 2014) and Mercer Subglacial Lake (accessed on 26 December. 2018; Priscu et al., 2021)—are located beneath Whillans Ice Plain. Those provide the only in situ constraints on subglacial water-column thickness, lake-water chemistry (Christner et al., 2014; Davis et al., 2023), and lakebed sediments (Hodson et al., 2016; Siegfried et al., 2023) available across the continent.

2.2. Radio-Echo Sounding Survey

During December 2013 and January 2014, the Center for Remote Sensing and Integrated Systems (CReSIS) surveyed the Gould and Siple coasts, West Antarctica, with their Multi-channel Coherent Radar Depth Sounder 4 (MCoRDS4) (Gogineni et al., 2014) mounted on a Basler BT-67 airframe. An eight-element antenna-array was mounted in the cross-track direction under the plane, and operated across a 150–450 MHz frequency band. This survey included multiple flights over Whillans Ice Plain and directly over several active subglacial lakes.

2.3. Bed-Echo Power Processing

The radar power returned from a bed echo is the result of a number of transmit and propagation effects, each of which must be considered before interpreting bed characteristics. Following Matsuoka et al. (2012), measured bed-echo power can be described as,

$$[P] = [S] - [G] + [R] - [L] - [B] \tag{1}$$

where S represents instrumental factors (e.g., transmit power and antenna gain), G represents geometric-spreading losses due to spherical wave-front propagation, R represents bed reflectivity (the property in which we are interested), L represents attenuation losses, B represents birefringence losses, and [ ] indicates that all values are on a decibel scale (i.e., $[x] = 10\log_{10}x$). Losses due to geometric spreading are

$$[G] = 2\left[2\left(h + d/\sqrt{\epsilon_r}\,\right)\right] \tag{2}$$

where h is aircraft survey height, d is ice thickness, and ϵ_r is the relative permittivity of ice (3.15). Losses due to attenuation through the ice column are

$$[L] = 2d < N > \tag{3}$$

where $<N>$ is the one-way, depth-averaged attenuation rate. To first order, $<N>$ varies with ice temperature and impurity content (Moore & Fujita, 1993), both of which are highly uncertain across most of the Antarctic ice sheet. Therefore, prior studies have either assumed $<N>$ is uniform across a survey area (Catania et al., 2003), calculated empirically based on the relationship between return power and ice thickness (e.g., Hills, Christianson, Jacobel, et al., 2022; Matsuoka, Morse, & Raymond, 2010; Schroeder, Seroussi, et al., 2016), or forward calculated using an ice-sheet model (e.g., Chu et al., 2018; Matsuoka et al., 2012).

Although large-scale analysis of an entire basin requires detailed accounting of each correction to determine an unbiased estimate of bed reflectivity, we are investigating changes in bed reflectivity across relatively short length-scales (10s of km), and thus corrections that vary over larger spatial scales (i.e., birefringence) are assumed to be uniform in our analysis. We also assume that transmit strength and system losses did not change within an individual mission (i.e., one day's flight). To correct for any differences between missions, we calculated per-mission crossover differences following Chu et al. (2021) and apply this correction, M, to level the data across the nine flight days. Because of the limited ice-thickness variation across our survey area, we were not able to empirically determine a value for $<N>$. Previous work with lower frequency RES on Whillans Ice Plain (e.g., Christianson et al., 2012; MacGregor et al., 2011) suggested a one-way, depth-averaged attenuation rate between 10 and 15 dB km^{-1}. We used $<N> = 10$ dB km^{-1} and performed a sensitivity test using values between 5 and 15 dB km^{-1} to assess how much $<N>$ impacts our result (Text S1 in Supporting Information S1).

We used single-channel RES images (i.e., radargrams) processed using the CReSIS toolbox with standard synthetic aperture radar focusing (Open Polar Radar, 2023). The provided bed picks are typically near, though not precisely at, the bed. Therefore, we re-defined the bed location and power using the range bin with maximum power within 100 m of the product-provided bed pick. We calculated ice thickness based on the difference between our bed pick and the product-provided surface pick. Finally, we combined Equations 1–3 to solve for relative bed-reflectivity,

$$[R_{rel}] = [P] + 2\left[2\left(h + d/\sqrt{\epsilon_r}\,\right)\right] + 2d < N > + M \tag{4}$$

now adding the crossover corrections, M, ϵ_r and again assuming that $[S]$ and $[B]$ are negligible (processing steps shown in Figures S1–S12 of Supporting Information S1).

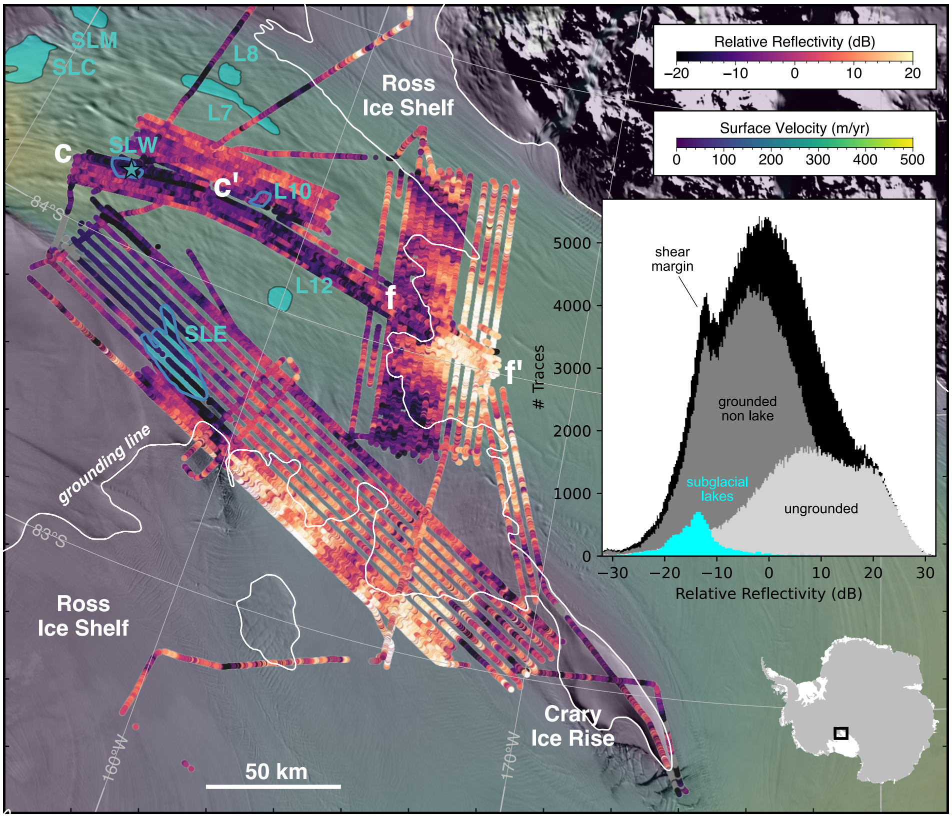

Figure 1. Map of the radio-echo sounding survey at Whillans ice plain with estimated relative reflectivity for each measured trace. Inset map shows the survey location within Antarctica. Inset histogram shows estimated relative reflectivity for: all traces (black); traces at the ice shelf (light gray); traces at grounded ice outside of subglacial lakes (dark gray); and traces at subglacial lakes (cyan). Lake histogram bins are twice as large as others for visibility at this scale. Reference data sets include lake outlines (Siegfried & Fricker, 2018), grounding line (Depoorter et al., 2013), and MODIS imagery (Scambos et al., 2007). Text indicates locations of subglacial lakes, two flight tracks highlighted in Figure 2, and other prominent glaciological features in the region.

3. Results

Our estimated relative bed reflectivity across Whillans Ice Plain spans ∼60 dB or six orders of magnitude (Figure 1). The ungrounded ice shelf is relatively bright compared to grounded ice, on average 10.8 dB higher relative reflectivity, as would be expected given the large dielectric contrast between seawater and ice compared to sediment and ice (∼5–29 dB; Christianson et al., 2016; Peters et al., 2005). In contrast, the four active subglacial lakes covered by this survey, SLW, Engelhardt Subglacial Lake (SLE), Lake 7, and Lake 10 have 10.4 dB lower mean relative reflectivity compared to off-lake grounded regions. This result is surprising given that reflections from liquid water should be 3–27 dB brighter than surrounding thawed or frozen till (Christianson et al., 2016; Peters et al., 2005), suggesting ∼13–37 dB net power suppression.

Three regions within Whillans Ice Plain have particularly dim bed-echoes, the northern ice-stream shear margin (Figures S16–S18 in Supporting Information S1), south of Lake 7 (Figures S19–S20 in Supporting Information S1), and down-glacier of SLW (including Lake 10). We emphasize this area between SLW and the grounding zone (Figure 2) which numerous studies have shown to be hydraulically active (e.g., Fricker & Scambos, 2009; Siegfried & Fricker, 2018). A variety of basal features are resolved in the radargrams (Figure 3). The subglacial lakes, including SLW, have relatively dim and diffuse bed echoes. Outside of known lake areas, we observe both specular and diffuse bed echoes. Unlike RES surveys elsewhere in Antarctica, we commonly observe volume scattering in the ∼10s of m above the bed echo (distinct from basal crevassing).

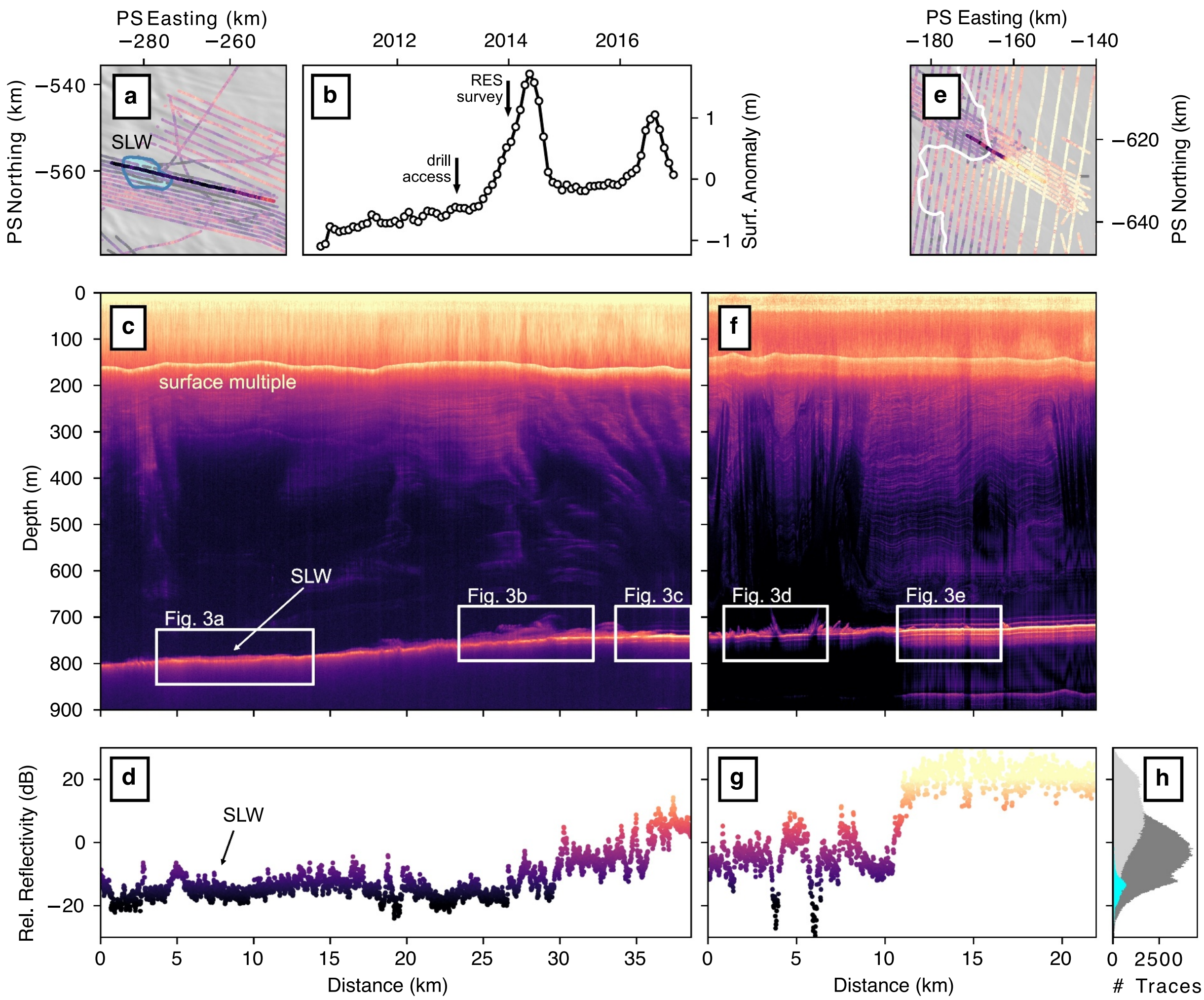

Figure 2. Observations at and down-glacier of SLW. (a) Map of SLW area, with estimated relative reflectivity for radio-echo sounding (RES) flight tracks. (b) SLW surface-elevation time series as an anomaly relative to surrounding ice (Siegfried & Fricker, 2018). (c) Radargram corresponding to selected flight track in (a). Note that variations in the reflected power of englacial layers (above the bed) are generally thought to be associated with layer dip and destructive interference (Castelletti et al., 2019; Holschuh et al., 2014). (d) Estimated relative bed reflectivity for the radargram in (c). (e–g) Map, radargram, and reflectivity as in a, c, and d but for the grounding-zone region down-glacier of SLW. (h) Estimated relative-reflectivity histogram for RES traces (colors as described in Figure 1).

4. Discussion

The RES survey we analyzed here captures four active subglacial lakes in three distinct catchments of Whillans Ice Plain. In all cases, the timing of the 2013–14 survey corresponds to a mean surface-height anomaly above low stand (Siegfried & Fricker, 2018). We can therefore refute one hypothesis for disagreement between observational techniques: that the lake was drained at the time of survey (e.g., Langley et al., 2011; Wright & Siegert, 2012; Welch et al., 2009). Still, if lake depth were thin relative to the radar wavelength, interference could suppress reflected power (e.g., Christianson et al., 2012; Siegert et al., 2014). Specifically for SLW, direct sampling 10 months before RES surveying showed a standing water column of 1.6–2.2 m (Tulaczyk et al., 2014); a continuous GPS station ~700 m away from the borehole at the time of drilling showed that surface height increased by 1.9 m between direct access and radar surveying (updated from Tulaczyk et al., 2014), suggesting a water column >3.5 m at the time of the survey. For the radar frequency here, $f_c = 300$ MHz, and for the measured lake conductivity (720 μS cm^{-1}; Christner et al., 2014) the skin depth is 0.11 m, so there is no destructive interference from any lake bottom reflection (Dowdeswell & Evans, 2004). Others have hypothesized that the lake area may be too small (e.g., Wright et al., 2014), but here the lakes are substantially larger than the diameter of the first Fresnel zone (~25 m). Thus, we propose that the ~13–37 dB suppression in estimated relative

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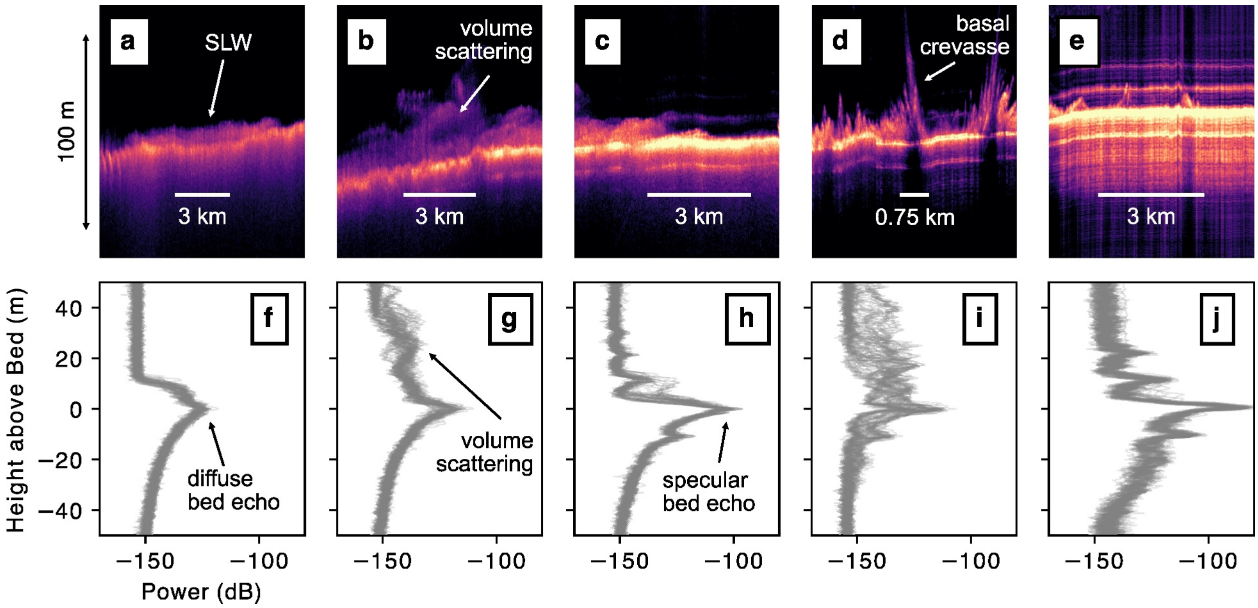

Figure 3. Representative basal features from radargrams. (a–e) Insets of radargrams for features of interest (inset locations shown in Figure 2). (f–j) Observed power profiles of traces spanned by the white line in a–e. (a) & (f) Diffuse bed-echo from SLW. (b) & (g) Volume scattering down-glacier of SLW. (c) & (h) Specular bed-echo down-glacier of SLW. (d) & (i) Grounding-zone basal crevassing. (e) & (j) Bright ice-shelf reflection.

reflectivity over these lakes compared to their surroundings is associated with some physical properties of the ice, water, or interface at an active subglacial lake.

After correcting for geometric spreading and discarding any consideration of birefringence over these short spatial scales, there are two mechanisms that can drive suppression in measured bed-echo power, scattering and attenuation. We suggest five glaciological processes which drive scattering and/or attenuation as hypotheses for the cause of lower estimated relative reflectivity compared to those observed in non-active lakes or calculated from theoretical reflection coefficients:

1. Roughness or anisotropy at the ice-bed interface causes (possibly directional) surface scattering losses;
2. Roughness at the ice-air interface causes surface scattering losses;
3. Entrained sediment in basal ice causes volume scattering losses;
4. Warm basal ice causes attenuation losses;
5. Entrained liquid water in pores, fractures, or along grain boundaries within basal ice causes both volume scattering and attenuation losses.

Below, we test these hypotheses with models of radar-wave propagation, grouped into scattering and attenuation effects (Figure 4).

4.1. Scattering

When the radar wave encounters a contrast in the medium permittivity, its propagation is disrupted and energy scatters. In a sense, this phenomena is helpful because some portion of that scattered energy returns to the receiver, but here we want to account for power losses due to scattering. First, consider surface scattering where the spatial extent of permittivity contrast is much larger than the propagating wavefront (e.g., the air-ice or ice-bed interfaces). Following Peters et al. (2005), there is a reduction in reflected power by a scattering surface, defined as

$$\rho_{surface} = e^{-g^2} I_0^2\left(\frac{g^2}{2}\right) \tag{5}$$

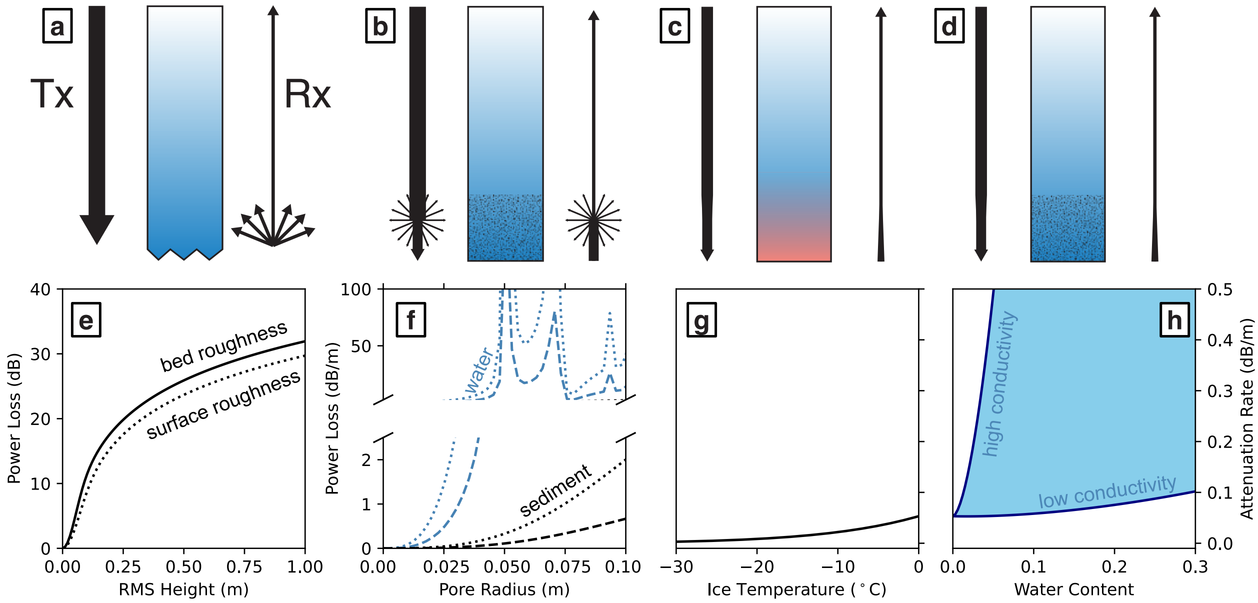

Figure 4. Physical mechanisms for suppression of radio-echo sounding bed-echo power, including illustrations (a–d) and a representative range over plausible glaciological conditions (e–h). (a) & (e) Surface scattering by either the ice-air or ice-bed interface. (b) & (f) Volume scattering by entrained water or sediment for both high (dotted) and low (dashed) porosity. (c) & (g) Attenuation by warm basal ice. (d) & (h) Attenuation by entrained water for both high and low conductivity.

where and I_0 is the zeroth order modified Bessel function of the first kind, and g is a root-mean square (RMS) phase variation at the scattering interface which is dependent on the interface shape relative to the radar wavelength

$$g = \frac{4\pi S}{\lambda} \tag{6}$$

where S is RMS height of the scattering interface and λ is wavelength. Considering the lid of a subglacial lake as the scattering interface in this context, power suppression at the lake would be observed if the lake lid were rougher than the ice-bed interface outside the lake. To the contrary, lakes are commonly identified as anomalously smooth subglacial interfaces (Carter et al., 2007), and Christianson et al. (2012) found no significant differences in the power spectrum of basal roughness within and outside of SLW. Alternatively, basal roughness is commonly anisotropic, aligned with the ice flow direction (Cooper et al., 2019). Here though, we observed power suppression for both along (e.g., SLW) and across-flow (Lake 7) flight tracks. We also note that prior studies which demonstrated power suppression at active subglacial lakes have been both along and across flow (e.g., Wright et al., 2014). We therefore reject hypothesis 1.

With propagation through the scattering interface (e.g., at the air-ice interface) a term must be added to Equation 6 for refraction at the interface (Schroeder, Grima, & Blankenship, 2016), now

$$g = \frac{4\pi S}{\lambda}\left(\sqrt{\epsilon_r} - 1\right) \tag{7}$$

As for the ice-bed interface, there is no observed reduction in ice-surface roughness above these lakes, neither in satellite laser altimetry (e.g., Fricker & Scambos, 2009) nor ground surveys (e.g., Christianson et al., 2012). We can therefore reject hypothesis 2 as well.

Next, we consider volume scattering where the medium through which the wave propagates contains many small particles or inclusions of different permittivity than the medium itself. In our case, volume scattering can be caused by air, water, or sediment in the ice column. Based on Mie scattering theory (Aglyamov et al., 2017; Eluszkiewicz, 2004), one-way scattering losses are

$$\rho_{volume} = e^{-\int Q \frac{\phi}{4/3r} dz} \tag{8}$$

where $Q = Q(r, \epsilon_\phi, \epsilon_r)$ is scattering efficiency calculated numerically (Prahl, 2023), r is the representative pore radius, ϕ is volume fraction or porosity, and ϵ_ϕ is permittivity of the pore material.

Near the ice surface, volume scattering may be caused by air-filled crevasses, even those buried beneath the ice surface. Within our survey, crevasses are most prevalent at the northern shear margin of Whillans Ice Stream (Fried et al., 2014), near SLE, and it is in fact the case that near-surface scattering is observed in those radargrams. However, that observed scattering extends beyond the lake area and is not directly coincident with the suppressed bed-echo power.

An alternative case for volume scattering as a cause for our observed power losses is more plausibly associated with active subglacial lakes themselves. That is, that hydraulic activity at the ice-bed interface causes entrainment of water and/or sediment into basal ice. Depending on the size of particles/pores, even a relatively small thickness (∼meters) of entrained material could scatter enough to cause observed power losses (Figure 4f). Our observations of volume scattering (e.g., Figures 3b and 3g) extend 10s of m above the bed reflector in some cases, easily enough to suppress bed-echo power, even for small entrained particles (∼cm). We cannot reject hypotheses 3 or 5 on this basis and reserve further discussion for Section 4.3.

4.2. Attenuation

In addition to scattering, some energy is attenuated from the radar wave and absorbed into the medium. In Section 2.3 we assumed a constant attenuation rate. Here, we consider possible variations in attenuation which could cause suppression of measured bed-echo power between traces. In ice, radar attenuation is primarily by electrical conduction (Corr et al., 1993; MacGregor et al., 2007; Moore & Fujita, 1993), so we represent one-way depth-dependent attenuative losses by

$$\rho_{attenuation} = e^{-\int \frac{\sigma}{\epsilon_0 \sqrt{\epsilon_r} c} dz} \tag{9}$$

assuming that electrical conductivity, σ, can change from top to bottom of the ice column. In an alpine glacier or for planetary ices there can be strong gradients in ice impurity and associated conductivity, possibly manifesting as a radar "fog" (Whitten & Campbell, 2018) which could obscure the bed echo. Earth's ice sheets have more gradual impurity gradients, with changes over much larger spatial scales (100s of km; Matsuoka, MacGregor, & Pattyn, 2010) than those considered here (10s of km across the lakes). Therefore, prior to the considerations of water entrainment in basal ice given below, we first use a pure-ice conductivity which is only dependent on temperature, T:

$$\sigma_i = \sigma_{i0} e^{\frac{E}{k}\left(\frac{1}{T_r} - \frac{1}{T}\right)} \tag{10}$$

where σ_{i0} is conductivity at the reference temperature, T_r, E is an activation energy, and k is Boltzmann's constant. As a crude overestimate of warming, assume that ice upstream of the lake is uniformly at the air temperature of −25°C. Then, the lake acts as an energy source, instantly warming the bed to 0°C and slowly warming the rest of the ice column. This scenario has an analytical solution with the complex error function, $\mathcal{E}$,

$$T(z,t) = \Delta T \mathcal{E}\left(\frac{z}{2\sqrt{\alpha t}}\right) \tag{11}$$

where ΔT is the magnitude of temperature change, 25°C, α is thermal diffusivity of ice, and t is time since basal warming started. Then, total attenuation in pure ice, from transmit to receive, is calculated with Equations 9–11, and integrating both ways through the ice column. For a change in bed-echo power of the magnitude we observe, >13 dB, to be associated with ice temperature would take ∼1,200 yrs of warming from the lake, even for this overly conservative estimate. At the ∼300 m/yr surface velocities, that is an advected distance of 400 km, 1–2 orders of magnitude longer than the scale of subglacial lakes in the surveyed area. In short, warming of basal ice is much too slow to cause the power loss we observe at these active subglacial lakes, and we can reject hypothesis 4.

Finally, consider water entrained in basal ice, as in the volume-scattering scenario above, which may be expected to vary over the spatial scale of a subglacial lake. Unlike in the consideration of scattering, the size of individual water pores is inconsequential because it is the electrical conduction in the water (not scattering) which causes loss of power. Following work describing power loss in firn aquifers (Chu et al., 2018; Kendrick et al., 2018) we define a mixed-media conductivity,

$$\sigma_{mixed} = \sigma_w(\phi - \phi_c)^\gamma + \sigma_i(1 - \phi) \tag{12}$$

where σ_w is source-water conductivity, ϕ is again porosity, ϕ_c is associated with connectivity between pores, and γ is an empirical parameter that controls the increase in conductivity due to material porosity. As in previous studies, we assume $\phi_c = 0$ and $\gamma = 1.67$ (Geldsetzer et al., 2009; Kendrick et al., 2018). We vary σ_w between the value of melted glacier ice at SLW (5.3 μS cm^{-1}) and SLW water (720 μS cm^{-1}; Christner et al., 2014) for porosities between 0 and 0.3. As for the volume-scattering case, even a relatively small thickness of entrained liquid water in basal ice (several meters) could attenuate sufficient radar power to cause the observed power loss, especially in the higher-conductivity SLW-water case.

4.3. Interpretation

Based on our analyses above, we argue that suppression of observed bed-echo power at active subglacial lakes is associated with entrained particles within basal ice. Since we have direct observations of volume scattering 10s of m above the basal interface (e.g., Figures 3b and 3g), similar to what has been hypothesized as sediment/debris in prior studies (Winter et al., 2019), we argue that process must be at least partially effective in bed-echo suppression. In cases with sufficient entrained material, the received radar power from volume scattering could be greater than from the ice-bed interface itself, or even obscure the ice-bed interface completely (Franke et al., 2023). If the entrained material is electrically conductive (e.g., SLW water), then attenuation is an additional factor. Based on the higher conductivity, and that lakes are inherently a water feature, we argue that water entrainment is a more plausible hypothesis than sediment. Water content in the basal ice is then important for understanding basal ice mechanics as well (Adams et al., 2021).

Since bed-echo power suppression is observed at active lakes only, there must be some unique physical process associated with lake activity (e.g., over pressurization or turbulent mixing during a drainage event) which forces water into basal ice. Non-active lakes, on the other hand, are slowly and consistently melting/freezing (MacGregor et al., 2009), with no turbulent fill/drain events, and are therefore observable with RES. Understanding differences between these two lake regimes beneath Earth's ice sheets and how those lead to different RES signatures will add fidelity to future planetary RES investigations (Roberts et al., 2023) as has already been shown by the debated nature of bright RES reflections from beneath Mars' southern ice cap (Schroeder & Steinbrügge, 2021). For those investigations, additional RES signatures should be considered; for example, reflection specularity (Schroeder et al., 2013), spatial variability in reflectivity and wet-to-dry transitions (Jordan et al., 2018), or statistical techniques which have primarily been used to study roughness of the ice-air interface (Grima et al., 2012, 2014).

5. Conclusion

In this study, we used previously-collected airborne RES data from Whillans Ice Plain to demonstrate that bed echoes from active subglacial lakes are 13–37 dB weaker than expected for a simple ice-water interface as compared to surrounding thawed or frozen till. Contrast this with the non-active subglacial lakes which have been routinely identified by their anomalously bright bed echo (Wright & Siegert, 2012). We formulated and tested five hypotheses for physical mechanisms which could suppress bed-echo power in the way we observe. We found that two related hypotheses are plausible, and argue that one is particularly likely. That is, that lake fill/drain activity and pressure-induced entrainment of water into basal ice causes both volume scattering and attenuation. We argue that these processes are unique to active subglacial lakes and that the hypothesized water content identified in ice 10s of m above the bed interface is important for understanding basal ice mechanics.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The processed RES data are openly available from CReSIS (2024) . Those data were generated with support from the University of Kansas, NASA Operation IceBridge grant NNX16AH54 G, NSF grants ACI-1443054, OPP-1739003, and IIS-1838230, Lilly Endowment Incorporated, and Indiana METACyt Initiative. Subglacial lake outlines and elevation time series were originally published and made available by Siegfried (2021). Computational notebooks to reproduce the derivative RES products and figures from this article are available at https://github.com/benhills/subglacial-lake-radar-power with a frozen version (Hills et al., 2024) as well as the derivative data products (Hills & Siegfried, 2024).

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