Cold-based debris-covered glaciers: Evaluating their potential as climate archives through studies of ground-penetrating radar and surface morphology

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Abstract We describe the morphology and internal structure of Mullins and Friedman Glaciers, two cold-based, debris-covered alpine glaciers that occur in neighboring valleys in the McMurdo Dry Valleys, Antarctica. Both glaciers are overlain by a single, dry supraglacial debris layer (8–75 cm thick); each mantling debris layer is marked with near-identical patterns of arcuate ridges and steps, as well as corresponding changes in bulk grain size, meter-scale surface topography, and thermal contraction crack polygons. Results from 24 km of ground-penetrating radar data show that the ice within the uppermost 1–2 km of Mullins and Friedman Glaciers is essentially free of englacial debris (<1% by volume) but thereafter is interspersed with bands of englacial debris (each <3 m thick and dipping up glacier) that intersect the ice surface at all major surface ridges and steps. The similarity in number and pattern of englacial debris bands and corresponding surface ridges and steps across both glaciers, along with model results and observations that call for negligible basal entrainment, is best explained by episodic environmental change at valley headwalls. Our working hypothesis is that layers of englacial debris originate as supraglacial lags that form in ice-accumulation areas during times of reduced net ice accumulation; following renewed net ice accumulation, these lags are subsequently buried by snow and ice, flow englacially, and intersect the ice surface to impart distinctive changes in the texture of supraglacial debris and topographic relief. The implication is that the englacial structure and surface morphology of these cold-based, debris-covered glaciers preserves a consistent record of climate and environmental change.

1. Introduction

Debris-covered viscous flow features that are cored with interstitial or massive buried ice are conspicuous, yet enigmatic landforms in glacial and periglacial environments. Current research aims to understand their origin(s) [Clark et al., 1998; Humlum, 1996; Martin and Whalley, 1987; Potter et al., 1998], ice-flow dynamics [Arenson et al., 2002; Haeberli et al., 2006, 1998; Kääb et al., 2003; Kääb and Weber, 2004], erosive and debris-transport capabilities [Barsch and Jakob, 1998; Haeberli, 1985; Humlum, 2000], and paleoclimate significance [Clark et al., 1996; Fitzpatrick et al., 1995; Frauenfelder and Kääb, 2000; Humlum, 1998; Kääb and Weber, 2004; Konrad et al., 1999; Steig et al., 1998]. In most cases, the layer of surface debris that caps these features tends to reduce ablation of underlying ice [Konrad and Humphrey, 2000; Kowalewski et al., 2006, 2011; Lundstrom et al., 1993; Mihalcea et al., 2006; Potter, 1972] and permits long-term survival of subsurface ice [Clark et al., 1998], possibly for more than 8 Ma [Sugden et al., 1995b].

The specific origin of ice in many of these features is largely unknown and can vary over time. For example, ice may originate as segregation ice, from the burial and compaction of snowbanks, from the densification of snow and firm as for typical glaciers, or from other sources [Burger et al., 1999; Clark et al., 1998; Giardino et al., 1987; Haeberli and Vonder Mühll, 1996; Johnson, 1984; Potter et al., 1998; Shroder et al., 2000]. Further, the long duration over which many of these features seem to form [Clark et al., 1998, 1996; Fitzpatrick et al., 1995; Potter et al., 1998] suggests that ice origins may evolve over time, yielding notable, and perhaps cyclic changes in the style of debris entrainment, ice-flow dynamics, and overall morphology [Ackert, 1998; Frauenfelder and Kääb, 2000; Haeberli et al., 1998; Haeberli and Vonder Mühll, 1996; Whalley and Azizi, 1994, 2003; Whalley and Martin, 1992].
In an effort to elucidate the climatic significance of some long-lived viscous flow features in Antarctica, we initiated in 1995 a suite of detailed geomorphological and geophysical studies of buried ice in interior regions of the McMurdo Dry Valleys (MDV) [Kowalewski et al., 2012, 2006, 2011; Marchant and Head, 2007; Marchant et al., 2002; Sugden et al., 1995b; Swanger and Marchant, 2007; Swanger et al., 2010]. Our initial results demonstrated that buried ice can survive for millions of years [Marchant et al., 2002; Sugden et al., 1995b], that sublimation and subsequent vapor diffusion through capping debris can be extremely slow on the order of <10^{-1} mm a^{-1} [Kowalewski et al., 2012, 2006, 2011; Marchant et al., 2002], and that the origin of buried ice deposits varies with local microclimate conditions [Marchant and Head, 2007; Marchant et al., 2013; Swanger et al., 2010]. Here we focus on the origin and internal ice-flow dynamics of two viscous flow features, previously mapped as cold-based debris-covered alpine glaciers: the Mullins valley and Friedman valley debris-covered glaciers (hereafter informally termed Mullins Glacier and Friedman Glacier). We selected Mullins and Friedman Glaciers for detailed study because (1) initial studies suggest that they are cored by glacier ice and thus represent the polar, end-member class of cold-based, debris-covered glaciers [Kowalewski et al., 2011]; (2) their proximity allows for direct analyses and comparison of ice-flow dynamics, debris entrainment, and landscape modification under shared climate conditions [Kowalewski et al., 2011; Shean et al., 2007; Shean and Marchant, 2010]; (3) they have been cited as terrestrial analogs of similar-appearing viscous flow features on Mars [Head et al., 2010; Levy et al., 2006; Marchant and Head, 2007; Marchant et al., 2010]; and (4) preliminary geophysical surveys suggest that the uppermost portion of Mullins Glacier, at least, contains a nonuniform distribution of englacial debris that may reflect environmental change [Shean and Marchant, 2010].

2. Background

2.1. Setting

Debris-covered glaciers in the MDV are relatively uncommon and appear restricted to inland regions where ice accumulates in the lee of steep cliffs, most notably beneath cliffs of Ferrar Dolerite. Mullins Glacier (77.89°S, 150.58°W, 1557 mean sea level (msl)) originates along rock cliffs of Ferrar Dolerite and Beacon Supergroup sandstones at the headwall of Mullins Valley, ~2100 m elevation. It flows along the southwest side of Vestal Ridge for ~3.5 km and then bends eastward into upper Beacon Valley, one of the major valleys that nearly bisect the Quartermain Mountains (Figures 1 and 2a). After traveling ~8 km, Mullins Glacier merges imperceptibly at 1350 m elevation with stagnant, Miocene-aged buried glacier ice derived from an ancient, southward flowing incursion of Taylor Glacier into central Beacon Valley [Kowalewski et al., 2011; Sugden et al., 1995b] (Figure 1). (Unless otherwise noted, reported distances along the glacier centerline are measured from the valley headwall).

Friedman Glacier (77.90°S, 160.51°W, 1540 msl), situated just to the west of Mullins glacier, likewise originates along steep cliffs of Ferrar Dolerite and Beacon Supergroup sandstones, flows northward from ~2050 m elevation alongside the westside of Rector Ridge (Figure 2b), and appears to terminate at a lobate front at 1500 m elevation and 3.8 km distant in upper Beacon Valley; buried ice from Friedman Glacier may, in fact, extend beyond this point [e.g., Kowalewski et al., 2011], but if so it is unreachable in hand-dug excavations and unrecognizable in ground-penetrating radar (GPR) data (see below). A small, unnamed debris-covered glacier (referred to herein as “DCG_03”) flows for 0.5 km alongside uppermost Friedman Glacier.

Data from interferometric synthetic aperture radar (InSAR) show that both glaciers display peak horizontal surface velocities near their respective valley headwalls of ~40 mm a^{-1} but thereafter decelerate to near 0 mm a^{-1} (within measurement error of ~±0.002 m a^{-1}) at ~3.6 km (Friedman) and 4.8 km (Mullins) [Rignot et al., 2002].

2.2. Ice Accumulation and Ablation

Ice accumulation is derived from a combination of windblown snow transported off the East Antarctic Ice Sheet (which is ultimately trapped in the lee of the steep cliffs of Ferrar Dolerite and Beacon sandstone that nearly encircle both Mullins and Friedman valleys) and from direct precipitation along valley headwalls (Figure 2). The relative contribution of each source is unknown, but the values have likely varied over time [Marchant and Head, 2007; modern-day regional snowfall is <50 mm water equivalent a^{-1} [Fountain et al., 2009]. A few isolated clasts of Ferrar dolerite, and to a lesser extent Beacon sandstone, dot the surface of the ice-accumulation area during summer months; some of these clasts lie along the distal end of linear
Figure 1. Part of the Landsat Image Mosaic of Antarctica showing the southern portion the McMurdo Dry Valleys (MDV). Centered at ~77°35′S, the MDV encompasses ~4000 km² of mostly ice-free terrain between the East Antarctic Polar Plateau (left) and the western Ross Sea (right). The white box in the lower left highlights the location of Mullins and Friedman valleys, both tributaries to Beacon Valley in the Quartermain Mountains and depicts the geographic location for Figures 3 and 4 and Figure S1 in the supporting information. The inset plots the location of the MDV (small blue box) on a sketch map of the Antarctic continent.

Figure 2. (a) Oblique aerial view of Mullins valley (view to the south) showing the accumulation area and a portion of the ablation area (Regions 1 and 2) of Mullins Glacier. (b) Oblique aerial view of Friedman valley (view also to the south) showing the accumulation area and a portion of the ablation area (Regions 1–3) of Friedman Glacier. Bedrock cliffs are Ferrar Dolerite (dark brown) and Beacon sandstones (light brown). (c) Large clast of Ferrar Dolerite (>35 cm in diameter) within the uppermost ablation area of Mullins Glacier (Region 1). The clast is perched on an ice pedestal at the center of a ~15 cm deep moat-like depression. (d) Similar clast as observed in Figure 2c, but with small gravel-and-cobble-sized clasts at the base of the central rock, and within the surrounding moat. (e) Dolerite clasts, <10 cm in diameter, embedded in ice within the uppermost ablation area of Mullins Glacier (Region 1); the clasts are covered with a very thin (<1 cm) layer of refrozen meltwater.
tracks in the snow, which can be traced upslope, tens-to-hundreds of meters, toward points of local detachment along isolated rock outcrops.

The transition to net ice ablation on both glaciers is irregular and spatially complex. In general, the transition appears to coincide with a subtle break in ice-surface slope at ~1600 m elevation. However, isolated blue ice ablation areas occur up to 2000 m elevation on both glaciers and attest to complex ablation patterns, most likely influenced by minor variations in solar radiation (shadowing, slope/aspect), and seasonal winds.

The ablation areas of Mullins and Friedman Glaciers both show abrupt transitions from exposed glacier ice (e.g., ice dotted with scattered and isolated rocks, and thin patchy gravels) to uniform coverage and burial beneath thin supraglacial debris (Figure 3). This transition coincides with the first major arcuate ridge on both glaciers and lies at ~1.5 km for Mullins Glacier and ~1.7 km for Friedman Glacier. Thereafter, supraglacial debris on both glaciers thickens down valley, increasing from ~7 cm to ~45 cm on Friedman glacier and from ~10 cm to a maximum of ~75 cm on Mullins Glacier (see section 4.4). As discussed in Kowalewski et al. [2011], the supraglacial debris is most likely sourced from rockfall at valley heads; additional input from valley sidewalls is unlikely because (1) large topographic depressions exist at the lateral margins of both Mullins [Kowalewski et al., 2011] and Friedman Glaciers, and (2) valley sidewalls are typically dry and do not appear to collect wind-blown snow in sufficient quantities to initiate detachment and rockfall (Figure 2) [Augustinus and Selby, 1990; Kowalewski et al., 2011]. Additional input from wind-blown sediment is likely minor. We base this assertion on the general paucity of wind-blown sediment in the Dry Valleys [e.g., Lancaster, 2002] and on the absence of suitable ice-free source areas to the south and southwest of Mullins and Friedman Valleys (e.g., in the direction of the strongest and most frequent winds in the region [Marchant et al., 2013]).

2.3. Cryoturbation

The cold and dry conditions limit active layer cryoturbation. However, for low-albedo dolerite rocks (albedo of 0.07 [Campbell and Claridge, 2006]), intense solar radiation may warm surfaces above 0°C, even though atmospheric temperatures remain well below 0°C (see also section 4.1). Wherever these warmed rocks come in direct contact with exposed glacier ice, as occurs in the uppermost ablation areas of both Mullins and
Friedman Glaciers, sublimation increases and minor melting may ensue; this meltwater may flow down glacier, in very small channels <2 mm deep, and ultimately refreeze in a thin layer of superposed ice just inward of the first topographic ridge on both glaciers (Figure 3; e.g., the frozen pond of Shean and Marchant [2010]; see also sections 4.4.1 and 4.4.2). Because the 0°C isotherm does not penetrate more than a few centimeters in the exposed rocks and debris [Kowalewski et al., 2006, 2011], glacier ice buried beneath at least ~10 cm of debris is perennially dry and remains frozen year round (e.g., Kowalewski et al. [2011]; see also section 4.1).

Given these environmental conditions, ablation beneath supraglacial debris ≥10 cm thick is accommodated solely by sublimation. Calculated rates of ice sublimation beneath Mullins debris are on the order of 0.01 to 0.001 mm of ice loss a\(^{-1}\) [Kowalewski et al., 2011] and soil gravimetric water content is ≤3% [Campbell and Claridge, 2006]; taken together, these factors imply negligible saturated active-layer cryoturbation, e.g., negligible lateral and vertical displacement of rocks and debris from alternate freezing and thawing of ice. This assertion is supported by (1) the finding of coherent signals in repeat synthetic aperture radar interferometry, which suggest negligible independent movement of surface boulders (at least over the 3 year measurement period, 1996–1999) [Rignot et al., 2002], and (2) the preservation of millimeter-scale microstratigraphy and in situ volcanic ash fall [Marchant et al., 1996, 2002, 2013], which suggests insignificant sediment sorting and involution over considerable time frames.

An important corollary that arises from these observations is that the surface morphology and grain-size characteristics of supraglacial debris on both Mullins and Friedman Glaciers, as well as the concentration and grain-size characteristics of shallow englacial debris (e.g., debris observed at and just below the buried ice surface), reflect the intrinsic properties of the glaciers themselves, rather than postdepositional modification and alteration via saturated, active layer cryoturbation (see also Kowalewski et al. [2011]). None of the clasts examined on or within Mullins and Friedman Glaciers exhibit glacial striations, facets, or polish.

### 2.4. Sensitivity to Environmental Change

For a typical clean-ice alpine glacier, surface albedo decreases and atmospheric temperature increases down-glacier, resulting in increased ablation towards the toe. Mullins and Friedman Glaciers adhere to this general increasing ablation trend with distance down-glacier, but only in their uppermost portions. Once supraglacial debris thickens beyond a critical threshold, ablation is reduced [e.g., Ostrem, 1959]. For Mullins and Friedman Glaciers, ablation is highest in the uppermost portion of the glacier and then decreases dramatically beneath a thickening cover of supraglacial debris in the down-ice direction (see also section 4.4; [Kowalewski et al., 2011]). This spatial pattern of ablation is characteristic of many debris-covered glaciers [Ackert, 1998; Konrad et al., 1999; Scherler et al., 2011]. For Mullins Glacier, the ablation rate decreases abruptly by over 2 orders of magnitude from a measured rate of ~6 cm a\(^{-1}\) for mostly exposed ice in the uppermost ablation area (see section 4.3) to a rate of ~0.5 mm a\(^{-1}\) across the first major topographic ridge (arcuate surface discontinuity (ASD)-1, see below and Figure 1), where surface debris abruptly thickens from ≤2 cm to 7–10 cm [Kowalewski et al., 2011]. The net effect is a distinct increase in sensitivity to climate change being focused on a relatively small area of mostly exposed ice directly down valley from the present accumulation area [Konrad et al., 1999]. Thus, any perturbation in environmental conditions that reduces net accumulation (ameliorating atmospheric temperatures, wind patterns less favorable to snow deposition, and/or overall precipitation/ice accumulation reduction) will result in disproportionate ice loss in the exposed uppermost ablation area, yielding low-elevation, “spoon-shaped” ablation hollows that merge down valley with elevated remnants of protected (debris-covered) glacier ice [e.g., Marchant et al., 2013]. The development of such spoon-shaped hollows has been discussed for debris-covered ice flows outside Antarctica [Ackert, 1998; Johnson and Lacasse, 1988; Potter, 1972; Whalley, 1979], and these features have been implicated as indicators of local environmental change. Within the McMurdo Dry Valleys, the features are described as “beheaded debris-covered glaciers” [Marchant and Head, 2007], one of which occurs in association with remnant, debris-covered glacier ice along the western margin of upper Beacon Valley [e.g., Marchant et al., 2013, Figure 7].

### 3. Methods

#### 3.1. Environmental Data Collection

##### 3.1.1. Micrometeorological Automatic Weather Stations

Micrometeorological conditions were monitored for up to 6 years using automatic weather stations (AWS) at seven locations on Mullins and Friedman Glaciers (see Figure S1 in the supporting information). The suite of
corers; core recovery at all sites was ≤ drilling photographed, and georeferenced using high-resolution GPS. To protect the pristine environment, no deep cores (15
We collected 21 shallow ice cores along the central
3.3. Ice Core Drilling

environmental sensors (all Onset Computer Corporation “smart sensors”) varied among AWS (see Table 1) but included a combination of (1) soil temperature and soil moisture sensors typically arrayed in vertical profiles down to the buried ice surface; (2) downwelling shortwave solar radiation; (3) surface barometric pressure; (4) air temperature and relative humidity (RH) at ground level (5 cm) and mast height (2 m); and (5) mast height (2 m) wind speed and direction (for individual sensor specifications including range and accuracy, see Kowalewski et al. [2006]). Each station sampled environmental conditions every 3 min and logged averages to HOBO® microstation U21 data loggers [e.g., Kowalewski et al., 2006]. Temporal coverage varies between 1 and 8 years, depending on installation date and occasional sensor failures (Table 1).

3.1.2. Borehole Temperature Monitoring

In November 2008, we recorded downhole temperatures for borehole MCI-08-01 (see section 3.3) located ~1.8 km from the headwall of Mullins Valley (77.89438 S, 160.58324 E) at 5, 10, 15, and 20 m depths. A second thermistor string was deployed at borehole MCI-09-03 (4.6 km from Mullins headwall, 77.8787 S, 160.53614 E) in December 2009 to monitor temperatures at 10 m, 20 m, and 30 m depths. Both arrays were deployed for a minimum of 5 days duration.

3.1.3. Ablation-Stake Array

In 2010 and 2011 we installed an array of ablation stakes in the uppermost ablation areas of Mullins and Friedman Glaciers: 14 stakes for Mullins and 5 stakes for Friedman (Figure S1). Each stake was constructed from 3.175 cm diameter PVC pipe (capped and filled with snow) and installed via drilling a 3.175 cm diameter hole ~70 cm down into glacier ice. The heights of the stakes above the glacier surface have been measured annually since, with measurements referenced to the top of a board (15 cm × 20 cm) placed over the adjacent ice surface; the latter minimizes errors associated with localized ice loss from minor wind scour immediately adjacent to stakes.

3.2. Geomorphic Mapping and Sedimentological Analyses

Morphometric analyses relied on field mapping and study of remotely sensed data. The latter included analyses of high-resolution aerial photographs (U.S. Geological Survey (USGS) TMA3079V0296 and TMA3080V0276, November 1993), multispectral and panchromatic satellite imagery (1.5 m and 0.41 m resolution, respectively, GeoEye01, January 2004), and airborne LiDAR data (2 m resolution) [Schenk et al., 2004]. We collected detailed topographic measurements and sediment samples at 75 sampling localities along both glaciers; samples included supraglacial debris and subjacent glacier ice (Figure S1). Each ~2 kg sample (approximately ~1 L) of supraglacial debris was sieved at 16 mm in the field; the >16 mm fraction was analyzed for lithology, shape, and weathering characteristics [e.g., Swanger et al., 2011], and the smaller fraction was stored in plastic Whirl-Pak® bags for standard grain-size analyses at Boston University [e.g., Kowalewski et al., 2011]. Unlike earlier sampling strategies that were designed to provide spatial coverage and detect general trends in grain-size characteristics and polygon morphology along Mullins and Friedman Glaciers [e.g., Kowalewski et al., 2011; Levy et al., 2006], our strategy here focused on obtaining numerous, closely spaced samples across mapped surface ridges and arcuate steps.

At the base of most soil excavations, we quarried 10–40 cm down into buried glacier ice to collect samples for δ18O and δD analyses, and total dissolved solids (TDS). We recorded the visible concentration of englacial debris, if present. Ice samples were double-bagged in Whirl-Pak® bags and kept frozen at ~20°C until analyzed. Isotopic analyses were conducted on an isotope ratio mass spectrometer at the Boston University Stable Isotope Laboratory. Analyses for δ18O were accomplished via CO2 equilibration; deuterium analyses were accomplished via pyrolysis using a GV Instruments Ltd. ChromeHD™ system. TDS were estimated from electrical conductivity (EC) measurements.

3.3. Ice Core Drilling

We collected 21 shallow ice cores along the central flowline of Mullins glacier (Figure S1). Seven relatively deep cores (15–30 m) were extracted using an electrically driven Koci Drill designed and manufactured by Ice Core Drilling Services for use in both clean and dirty ice glaciers [Green et al., 2007]. Fourteen shallow cores, each ≤ ~5 m, were obtained using Snow, Ice, and Permafrost Research Establishment and Kovacs Mark II hand corers; core recovery at all sites was >95%. Drill sites were cleared of supraglacial debris (if present), photographed, and georeferenced using high-resolution GPS. To protect the pristine environment, no drilling fluids were used. Drill heads on the KOCI drill were changed from ice cutters to rock-drilling heads whenever isolated rocks and/or thick (>10 cm wide) sand wedges were encountered [Green et al., 2007; Kowalewski et al., 2011]. Maximum penetration using the KOCI drill was ~30 m (at MCI-09-03); the primary
### Table 1. Micrometeorological Observations

<table>
<thead>
<tr>
<th>Physical</th>
<th>Friedman</th>
<th>Mullins</th>
</tr>
</thead>
<tbody>
<tr>
<td>Debris thickness (cm)&lt;sup&gt;b&lt;/sup&gt;</td>
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<td>0</td>
</tr>
<tr>
<td>Observation period (MM/YY)&lt;sup&gt;c&lt;/sup&gt;</td>
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<td>12/10–01/13</td>
</tr>
<tr>
<td>Location</td>
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<td>Longitude</td>
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<td>160.52429°E</td>
</tr>
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<td>Elevation above sea level (m)</td>
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</tr>
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<td>53 (48)</td>
</tr>
<tr>
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<td>Mean (m s&lt;sup&gt;−1&lt;/sup&gt;)</td>
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<td>Maximum (m s&lt;sup&gt;−1&lt;/sup&gt;)&lt;sup&gt;f&lt;/sup&gt;</td>
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<td></td>
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<sup>a</sup>No data indicated by "-". Statistical values are listed for both summer-only (defined as D-J-F) and annual (in parentheses) observations. NA, not applicable.<br><sup>b</sup>Both M_Met_01 and M_Met_02 are located in Region 1 in areas devoid of any surface debris layer.<br><sup>c</sup>The observation period for each station varied due to staggered initial installation dates and occasional data logger failure. The time periods listed here refer to the period for the entire station. All statistical valuations account for only the total time the station was in service; no specific adjustments are made to incorporate the seasonality of any outages. The observation intervals for the wind measurements are listed separately to reflect their higher volatility.<br><sup>d</sup>The total time the sensor recorded temperatures ≥0°C divided by the number of years in the observation record.<br><sup>e</sup>Anemometers sensors failed at a higher rate than all other sensors.<br><sup>f</sup>Because the anemometers failed during high-magnitude wind events, this value likely underestimates the true maximum velocity.<br><sup>g</sup>Measurements from 1 cm under surface of the debris layer.<br><sup>h</sup>Measurements at the base of the debris layer at the interface with massive ice (except for M_Met_01 which has no debris layer).
obstacle in drilling deeper was frictional heat (increasing the probability of drill entrapment downhole via the refreezing of meltwater) generated by cutting through rock and ice-rich sediment. The core data were used to help calibrate GPR velocity (see section 4.5) and ground-truth mapped englacial reflectors.

3.4. Ground-Penetrating Radar

We used a Geophysical Survey Systems, Inc. (GSSI) Subsurface Interface Radar 3000 ground-penetrating radar (GPR) system at multiple antenna frequencies (80 MHz, 200 MHz, and 400 MHz) to collect over ~24 km of common offset GPR data from Mullins and Friedman Glaciers (Figure 4).

Although radar investigations of ice-and-rock mixtures typically use low antenna center frequencies in the range of 15–50 MHz [Berthling et al., 2000; Degenhardt, 2009; Degenhardt and Giardino, 2003; Hausmann et al., 2007; Isaksen et al., 2000; Monnier et al., 2008], we follow Shean and Marchant [2010] and De Pascale et al. [2008] in selecting higher-frequency antennas in the range of 80–400 MHz that are capable of achieving both high spatial resolution and sufficient depth penetration. Due to the excellent transmission properties of the relatively clean and cold glacier ice, the 200 MHz and, in specific locations, the 400 MHz were successful in reaching depths of ~100 m. However, to achieve these depths, the 200 MHz and 400 MHz antennas required close physical contact with the buried ice surface, and thus, considerable effort was required to remove surface debris in order to expose buried glacier ice (or at least get to within a few centimeters of its surface). By contrast, the 80 MHz antenna was capable of achieving penetration of 100+ m without the necessity of removing surface debris. This capacity greatly enhanced our ability to maintain minimal environmental disturbance, enabling preservation of the unique morphology and stratigraphy of surface debris along GPR transects. Transects were collected along both longitudinal (down glacier) and transverse paths (Figure 4).

3.4.1. Survey Execution

These surveys were conducted in either distance mode, where a terrain-calibrated survey wheel triggered individual scans as a function of distance, or point mode where several scans were stacked at each prescribed location.
distance step. All long (>400 m) transects were collected using the unshielded Multiple Low Frequency (MLF) antenna in common offset mode configured for a center frequency of 80 MHz. Shorter GPR profiles collected with the 200 MHz and 400 MHz antennas were conducted on naturally exposed glacier ice, over glacier ice covered with snow, or where acceptable connection with the buried ice surface could be achieved with minimal removal of supraglacial debris. For the latter, all clasts > ~20 cm were cleared along the transect line, permitting close connection between the antenna base and the buried ice surface; typically a vertical separation between antenna and ice of <10 cm was achieved. For additional details, see Table S1 in the supporting information.

A Trimble 5700 Global Positioning System (GPS) receiver with a Zephyr geodetic antenna was used to collect high-resolution GPS (12 min minimum residence time) at each survey end point and continuous topographic GPS points along each survey line. After differential location correction utilizing data from a base station located ~7 km away on University Peak (77.86259°S, 160.75983°W; 2195.3 m), all position and elevation data displayed millimeter- to centimeter-scale accuracy. Points were imported into ArcGIS 10.0, and continuous elevation profiles were extracted from an airborne LiDAR digital elevation model (DEM) for Beacon Valley [Schenk et al., 2004]. Differences between the measured GPS and DEM elevations are minimal (mean = –0.33 m, 1σ = 0.39 m).

3.4.2. Vertical and Horizontal Resolution

The theoretical vertical resolution of a radar signal is calculated using vertical resolution = λ/4 = (V/F)/4 [Reynolds, 1997], where λ is the center wavelength of the antenna (m), V is the radar wave velocity of the medium (m s⁻¹), and F is the center frequency of the antenna (MHz). However, the actual vertical resolution will likely be significantly lower due to the progressive filtering of higher-frequency elements from the waveform [Hubbard and Glasser, 2005; Trabant, 1984], which may be particularly noticeable for debris-covered ice bodies [Berthling et al., 2000]. For this reason, we calculate a conservative estimate of vertical resolution = λ/3 for each of the antennas, yielding 0.15 m, 0.3 m, 0.75 m, for the 400 MHz, 200 MHz, and 80 MHz antennas, respectively, using a velocity of 0.168 m ns⁻¹ (see section 4.5 for velocity determination).

We follow the simplified approach of Welch et al. [1998] and use λ/2 as an approximate minimum size for general horizontal object resolution in migrated data. This approach results in working horizontal resolutions of 0.2 m, 0.4 m, and 1.1 m for the 400 MHz, 200 MHz, and 80 MHz antennas, respectively.

3.4.3. Postprocessing

GSSI Radan 6.7 and Radan 7.0 softwares were used in postprocessing GPR data. Processing steps common to almost all survey lines included the following: (1) horizontally appending individual data files comprising multipart survey lines; (2) static position correction to bring air-wave return to the top of the scan time window; (3) distance normalization using GPS and/or survey wheel data; (4) application of a finite infinite response filter with a 50/110 MHz, 120/220 MHz, or 340/480 MHz band-pass boxcar filter for the 80 MHz, 200 MHz, and 400 MHz survey data, respectively; and (5) gain adjustment. Additional processing steps, performed on a subset of survey lines, included (6) background removal, (7) migration, (8) Hilbert transform, and (9) topographic surface correction. Background removal is a spatial filtering process to remove constant artificial horizontal banding across the radargram that is caused by ringing between the source and receiver antennas and which may mask subtle “real” radar features. All of the 80 MHz survey data displayed mild to severe banding in the range 0–~ 200 ns, possibly caused by the relatively close spacing of the source and receiver antennas producing partial signal oversaturation [Hauck and Kneisel, 2008]. For these data, we applied a background removal spatial filter of length 71 scans, over the range window 21 ns to 250 ns. We then performed a simple two-dimensional Kirchoff migration assuming a mean velocity of 0.168 m ns⁻¹ (see section 4.5) to all survey data. Although migration at this velocity produced excellent hyperbola collapse of point reflectors in most profiles, we note the probable distortion of steeply dipping and complex continuous reflectors that may arise from the migration of radar data acquired in topographically complex environments [Lehmann and Green, 2000]. The large step size (0.7 m to 1.4 m) used in the 80 MHz surveys (relative to the signal wavelength) further reduced the efficacy of applying traditional migration. For this reason, data have been displayed in both migrated and unmigrated forms. For many of the migrated profiles, we performed an additional Hilbert magnitude transform [e.g., Yilmaz, 2001] to isolate the energy envelopes of individual scans [Arcone, 1996]. Finally, both the migrated and unmigrated data were topographically corrected based on 2 m elevation data extracted from a LiDAR DEM [Schenk et al., 2004] and confirmed with postprocessed kinematic GPS elevation data (section 3.4.3).
4. Results: Field Data

4.1. Automatic Weather Stations

Measured atmospheric conditions within the uppermost ablation areas of both Mullins and Friedman Glaciers (Region 1; Figure 3) are nearly identical and show persistent air temperatures well below 0°C (average of ~19°C) and low-average values of relative humidity (RH) near ~53% (Table 1). At the lowest elevation meteorological station (Figure S1), M_Met_02, located 4.2 km down valley from the headwall on Mullins Glacier (1376 m elevation), near-surface air temperatures (measured 5 cm above the ground surface) exceeded the melting point for an average of only 3.1 h each year. Temperatures at the buried ice surface at this site (at 25 cm depth) and elsewhere were consistently <0°C. Complete meteorological data are provided in Table 1.

4.2. Borehole Temperatures

Borehole temperatures show a general increase with depth, reaching ~23.2°C and ~21.5°C near the base of MCI-08-01 (at 20 m depth) and MCI-09-03 (at 25 m depth), respectively (Figure 5). The 15 m depth borehole temperature within both temperature strings is very close to the mean annual atmospheric temperature at both core sites. This is in agreement with the standard approximation of 10–15 m depth for the maximum penetration depth of the seasonal thermal wave [Paterson, 1994; Wade, 1945] and for a thermal system dominated by conduction only; it also supports the assumption that there are no englacial heat sources or sinks including melting/freezing or strain. The nonlinear concave downward shape of both borehole profiles at the shallowest depths most likely reflects the lingering influence of the seasonal wave at the time of sampling.

4.3. Ablation Measurements

Repeat measurements of the ablation-stake arrays on both glaciers show that ablation of exposed glacier ice in the uppermost ablation areas on Mullins and Friedman Glaciers average 5.56 and 6.86 cm a⁻¹, respectively (see Table S2 in the supporting information). Mullins stake 01, the only stake located on the secondary frozen meltwater pond (Figures 2, 3, and S1) [Shean and Marchant, 2010] (see also section 4.4.1.3), recorded a net increase of 2.5 cm between 2011 and 2013 (Table S2).

4.4. Ground Penetrating Radar, Ice Cores, and Surface Geomorphology: Organization of Data

Data from ground penetrating radar (GPR), shallow ice cores/soil excavations, and surface morphology reveal striking similarities across Mullins and Friedman Glaciers. These similarities, especially those observed in GPR, provide an informal basis for partitioning the ablation areas of Mullins and Friedman Glaciers into four distinct regions. Region 1 includes the uppermost ablation areas, whereas Regions 2–4 represent further subdivisions across the remaining, fully debris-covered portions of each glacier (Table 2 and Figure 3). Below, we present data from GPR transects, shallow ice cores and hand-dug ice excavations, and geomorphologic analyses of surface features for each region. For consistency, we provide data for Mullins Glacier first, followed by Friedman Glacier. Section 4.6 provides a short synthesis of key interpretations for all regions.
Table 2. Physical Characteristics of Mullins and Friedman Debris-Covered Glaciers

<table>
<thead>
<tr>
<th></th>
<th>Mullins Glacier</th>
<th>Friedman Glacier</th>
</tr>
</thead>
<tbody>
<tr>
<td>General</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Headwall start elevation (m)</td>
<td>2247</td>
<td>2227</td>
</tr>
<tr>
<td>Ablation area start (GPR start) elevation (m)</td>
<td>1652</td>
<td>1667</td>
</tr>
<tr>
<td>GPR end elevation (m)</td>
<td>1341</td>
<td>1400</td>
</tr>
<tr>
<td>Massive ice end elevation (m)</td>
<td>?</td>
<td>1413</td>
</tr>
<tr>
<td>Full glacier horizontal length (m)</td>
<td>&gt; 6000</td>
<td>3952</td>
</tr>
<tr>
<td>GPR coverage horizontal length (m)</td>
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<td>2900</td>
</tr>
<tr>
<td>Average slope (°) (not including HW)</td>
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<td>4.24</td>
</tr>
<tr>
<td>Nadir area, active flow only (km²)</td>
<td>3.39</td>
<td>2.59</td>
</tr>
<tr>
<td>Ice volume (not including HW) (m³)</td>
<td>1.0 – 1.6 × 10⁸</td>
<td>0.8 – 1.2 × 10⁸</td>
</tr>
<tr>
<td></td>
<td>ACA³</td>
<td>Region 1e</td>
</tr>
<tr>
<td></td>
<td>Mullins</td>
<td>Friedman</td>
</tr>
<tr>
<td>Start elevation (m)</td>
<td>2247</td>
<td>2227</td>
</tr>
<tr>
<td>End elevation (m)</td>
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<td>1667</td>
</tr>
<tr>
<td>Horizontal length (m)</td>
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<td>1400</td>
</tr>
<tr>
<td>Average slope (°)</td>
<td>4.32</td>
<td>4.24</td>
</tr>
<tr>
<td>Area (km²)</td>
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<td>0.408</td>
</tr>
<tr>
<td>Proportion of active flow (%)</td>
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<td>-</td>
</tr>
<tr>
<td>Ice volume (m³)</td>
<td>4.5–6.8</td>
<td>5.2–7.7</td>
</tr>
<tr>
<td>Average annual ablation rate (cm a⁻¹)</td>
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<td>-</td>
</tr>
<tr>
<td>Velocity (mm a⁻¹)</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Supraglacial debris coverage (%)</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Supraglacial debris thick. (cm)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Patterned ground coverage (%)</td>
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<td>-</td>
</tr>
<tr>
<td>Average polygon diameter (m)</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Average polygon trough depth (cm)</td>
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<td>-</td>
</tr>
<tr>
<td>Near surface englacial debris concentration (%)</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

*No data indicated by “-”.

**The total nadir surface area of the actively flowing portion of each glacier includes the headwall and accumulation zones. The actively flowing portion of each glacier, as determined from Rignot et al. [2002], includes the portion up to 5.0 km and 3.8 km on Mullins and Friedman Glaciers, respectively.

†Ice volume was determined by the integration of the local cross-sectional area along the desired distance. The cross-sectional area was estimated using the local width at the glacier surface and the estimated local depth from the GPR (see section 4.4). In agreement with the transverse GPR profiles, both rectangular and parabolic-shaped cross-sectional areas were used to calculate a range in the estimated ice volumes.

Ice volume determined from GPR as above (see note (c)) for regions specified.

*Supraglacial debris coverage (%) defined as the percentage of the total (nadir) area of the actively flowing portion of the glacier.

†Ablation rates in ACZ determined from stake data. Ablation rates estimates for all other regions have been estimated following Kowalewski et al. [2011].

Ice debris content estimated for a ~0.75 × 0.75 × 0.5 m³ (x × y × z) ice volume exposed and sampled at the base of each pits. Ice debris content values include both dispersed fine-grain debris and large grain-size debris. It should be noted that the maximum debris content values in Region 2 for both glaciers were very localized and obtained by sampling directly within an inclined debris layer (IDL) (see section 4.4).
4.4.1. Region 1: Mullins Glacier

4.4.1.1. GPR Data

Region 1 of Mullins Glacier extends from 0.6 to 1.5 km (Figure 4). The combined 200 MHz/80 MHz longitudinal radar transect along Mullins Glacier, $A_{m}-A_{m'}$, (Figure 6), is devoid of any major reflections for the first 600 m, except for a prominent basal reflector. This basal reflector descends sharply from a depth (two-way travel time) of ~675 ns at the beginning of the transection to a low point of ~1500 ns at ~1.1 km; thereafter, the basal reflector begins a slight incline toward the surface (Figure 4). Small and isolated hyperbolic reflectors are rare but locally present at all depths.

4.4.1.2. Ice-Core Data

Four shallow ice cores (maximum depth of 11 m for core MCI-11-02) obtained from Region 1 of Mullins Glacier all show clean, bubbly ice that is largely devoid of debris (Figure 7). Cores that penetrate the region of...
superposed ice in the distal end of Region 1 met refusal (additional penetration became impossible) at a shallow, horizontal rocky layer at ~80–90 cm. This rocky layer, visible at the lateral margins of the superposed ice, is well displayed in GPR profiles using a 200 MHz antenna [see Shean and Marchant, 2010].

4.4.1.3. Geomorphic Data

Boulders, cobbles, and gravels dotted across exposed ice in Region 1 cover from ~1 to 15% of the glacier surface; concentrations typically increase down valley (Figures 2 and 6). Where present, small patches of gravel attain a maximum thickness of ~1 to 2 cm. Isolated boulder-and-cobble-sized clasts are angular, and many display visible impact scars that likely reflect collisions during rockfall from the valley headwall (percussion marks [Kowalewski et al., 2011]). Overall, clasts show a subtle increase in slight surface staining (precursor to oxidation [Salvatore et al., 2013]), which becomes more obvious with increasing distance down valley.

The centimeter-scale microtopography of the ice surface in Region 1 is closely coupled to the distribution, size, and sorting of the overlying, scattered surface debris. The largest cobble-and-boulder-sized clasts of Ferrar Dolerite are typically perched on ice pedestals that lie at the center of shallow (15 cm) moat-like depressions (Figures 2c and 2d). These depressions are best developed on the up valley sides of rocks (e.g., upwind), with “tails” of ice and snow occurring in some places on the down valley sides. A sparse mixture of gravel-sized clasts typically collects at the base of many depressions; in places, these smaller clasts are lodged beneath the large, central rock, producing an unusual sorting pattern whereby large boulders and cobbles rest directly on small gravel-sized clasts (Figure 2d). Elsewhere, dolerite cobbles of Region 1 are found immediately below the ice surface, covered with a very thin (<1 cm) layer of refrozen meltwater (Figure 2e), or embedded to greater depths (up to ~20 cm), with a “tube” of refrozen meltwater connecting them to the ice surface (e.g., at the base of cryoconite holes [Fountain et al., 2004]). As noted above, the down valley
limit of Region 1 on Mullins Glacier displays a small area of superposed ice (12,490 m², 1.5–1.8 m deep) that in plan view appears as an oblong, frozen “pond” (Figures 3 and 6).

4.4.2. Region 1: Friedman Glacier

4.4.2.1. GPR Data

Region 1 of Friedman Glacier extends from 0.9 to 1.7 km. Beginning ~900 m from the headwall of Friedman Valley, the first 40 m of the combined 200/80 MHz longitudinal radar transect along Friedman Glacier, A⇑–Aʹ (Figure 4), appears clean and lacks major reflections apart from a prominent basal reflector (Figure 8). This basal reflector descends from a depth of 775 ns at the beginning of the transect and reaches a maximum depth of 1240 ns at 1250 m before rising gradually toward the surface. Dispersed hyperbolic point reflections and a strong horizontal reflection at 20 ns depth are apparent in the shallow 400 MHz profile (Figure 8).

Figure 8. Longitudinal transect A⇑–Aʹ for Friedman Glacier. (first panel) Vertical aerial photograph (USGS TMA 3080 V0276) showing transect location. (second panel) Unmigrated radar data. (third panel) Migrated data with a Hilbert magnitude transform applied. (fourth panel) Sketch of major radar reflectors and physical interpretation. The vertical depth scale for all panels is shown as two-way travel time (TWTT in ns) and as meters assuming an average radar travel time of 0.168 m ns⁻¹. Double-arrowed lines between the first panel and the second panel mark the location of identified arcuate surface discontinuities (ASD) and highlight the spatial correlation between ASD and inclined debris layers (IDL). Longitudinal distance along the transect line is shown as both the distance from the start of the transect and as the distance from the valley-headwall datum (see also Figure 4). Data for the first ~750 m were collected using a 200 MHz antenna in distance mode; thereafter, we used an 80 MHz antenna and collected data in point mode. The locations of Regions 1–4 are plotted on bottom axis.
4.4.2.2. Geomorphic Data
A mixture of boulders, cobbles, and gravels is sparsely distributed over 3–35% of the ice surface in Region 1 of Friedman Glacier; where present, small patches of gravel-sized clasts can locally achieve a maximum thickness of ~3 cm. Debris is most concentrated near the down valley end of the region. Clast characteristics and ice surface microtopography are nearly identical to those found on Mullins Glacier. Just as for the case of Mullins Glacier, the down valley limit of Region 1 displays a small area of superposed ice (14,880 m², 1.5–1.8 m deep) that appears in plan view as an oblong, frozen pond (Figures 3 and 8).

4.4.3. Region 1: Interpretation
The salient subsurface features within Region 1 of both Mullins and Friedman Glaciers include a strong basal reflector overlain by mostly homogeneous material that displays isolated hyperbolic reflectors. Along with the data from shallow ice cores, this suggests that the predominately homogeneous and reflector-poor areas represent clean glacier ice that is devoid of significant foliation or debris layers [e.g., Berthling et al., 2000]. The strong coherent negative-positive-negative polarities of the basal reflector most likely represent the intersection with valley beds. The smooth geometry of the basal reflections, as well as the sharp degradation in signal penetration below them, suggests glacially sculpted bedrock rather than loose, unconsolidated, and irregularly shaped subglacial debris/talus. The isolated hyperbolic reflections dotted throughout the reflector-poor areas, and especially apparent in the higher-frequency transects (e.g., 400 MHz), most likely represent individual/small groupings of clasts within the ice. This interpretation is supported by the high signal amplitude and negative-positive-negative polarity signature of these reflections [Arcone et al., 2002].

The microtopography of the ice surface in Region 1 is directly related to the uneven distribution of surface debris. The largest rocks alter proximal winds, producing up valley wind scoops and, in places, down valley ridges. Further, enhanced ablation along clast margins (but not at their base) gives rise to ice pedestals beneath the largest rocks [Doran et al., 2000; Hendy, 2000]. The inverse grading that places large cobbles and boulders on top of small gravel-sized clasts most probably arises from the eventual toppling of these “pedestal” rocks on gravel-sized clasts trapped in local depressions.

4.4.4. Region 2: Mullins Glacier
4.4.4.1. GPR Data
Region 2 of Mullins Glacier begins at the abrupt onset of the continuous supraglacial debris and extends from 1.5 to 4.1 km (Figures 3 and 6). The GPR data from this region are dominated by six distinct englacial reflectors that, in longitudinal transects, predominately dip up valley, and in transverse profile appear parabolic in shape (Figure 6). These reflectors are denoted as Mullins Inclined Reflectors 1, 2, 3, etc. (e.g., MIR 1 through MIR 6) and are described in detail below. Interspersed with these inclined reflectors are isolated hyperbolic reflectors, which are observed in all 80 MHz, 200 MHz, and 400 MHz transects. The basal reflector continues its trajectory from Region 1 but is increasingly diffuse with distance down valley (Figure 6); it eventually shoals to a depth of 190 ns at 3400 m along GPR transect $A_m-A_m'$ before deepening slightly to 310 ns at 4000 m.

MIR1 emerges at ~1050 m from a position slightly above the basal reflector in Region 1 and intersects the ice surface at 1600 m, at the apex of the first arcuate surface ridge and onset of continuous supraglacial cover (Figure 6); the angle of intersection of MIR 1 with the ice surface is ~30°. Close inspection of transect $A_m-A_m'$ reveals a distinct change in MIR 1 at ~1500 m, where the reflector deepens slightly before continuing upward toward the surface. Transverse profile $C_m'-C_m'$ (Figure 9) crosses this reflector and reveals its parabolic shape, nearly 500 m wide at the transect location; the reflector intersects the ice surface just beneath the lateral margins of the first surface ridge.

MIR 2 begins at ~1100 m and rises upward slightly before remaining nearly bed parallel for 300 m. Ultimately, MIR 2 intersects the ice surface at 2000 m, showing an intersection angle of ~35° (Figure 6). Analysis of a detailed portion of MIR 2 indicates that it is composed of a series of closely spaced, aligned hyperbolic reflections (Figure S3 in the supporting information). Data from transverse profile $C_m'-C_m'$ show that MIR 2 is parabolic in shape and that the deepest up valley portion of this reflector is nested beneath MIR 1 (Figure 9).

MIR 3 displays the most complex geometry of any of the identified inclined reflectors in Mullins glacier. The basal portion of this reflector is difficult to differentiate from MIR 2, but it appears to emerge just above the bed near ~1700 m; it intersects the ice surface at 2550 m, with an intersection angle of ~6° (Figures 6 and S3). The general upward trend of MIR 3 is interrupted by three steps where, at each step, the reflector remains nearly bed parallel for ~300–400 m (Figures 6 and S3). The last of these three bed-parallel steps occurs just beneath the
ice surface where the primary reflector splits into two reflectors: the primary reflector intersects the surface at 2500 m, and the secondary reflector remains below the surface for an additional ~50 m before intersecting the surface at 2550 m (Figure S3). MIR 3 is ~300 m wide where it intersects the ice surface at the location of transverse profile \( E_m - E_m' \) (Figure S4 in the supporting information).

MIR 4 is the least distinct of the major inclined reflectors. It emerges from near the bed at ~1700 m and remains nearly bed parallel for much of its length before ascending and intersecting the ice surface at ~2200–2300 m; the angle of intersection is ~17° (Figure 6). MIR 4 is 410 m wide where it intersects the surface along transverse profile \( F_m - F_m' \) (Figure S5 in the supporting information).

MIR 5 has a unique geometry departing significantly from the other major reflectors. Beginning at 2800 m, it takes the form of a large deformed U-shaped reflector reaching a depth of ~25 m at its lowest point (Figure 6). The down valley extension of MIR 5 intersects the surface at 3000 m with an intersection angle of ~35°. MIR 5 is 510 m wide where it intersects the surface along transverse profile \( F_m - F_m' \) (Figure S5).

MIR 6, the final inclined reflector, defines the boundary between Region 2 and Region 3 (Figures 3 and 6). It is unique in that it does not display a clear bed-parallel extension but instead emerges at a steep angle from near the bed at 3990 m and intersects the ice surface at 4020 m (intersection angle of ~26°) (Figure 6). The surface intersection of MIR 6 coincides with the down valley margin of a major ice lobe, showing ~15 m relief, which, in turn, roughly coincides with the initial descent of Mullins Glacier into upper Beacon Valley (Figures 3, 4, and 6). MIR 6 is ~250 m wide where it intersects the ice surface along transverse profile \( H_m - H_m' \) (Figure S6 in the supporting information).

**4.4.4.2. Ice Excavations and Ice Core Data**

Eleven shallow ice cores recovered from Region 2 of Mullins Glacier (all <24 m deep) show clean glacier ice interspersed with isolated, sand-and-gravel-sized debris < 1% by volume (Figure 7).

Fifty-five exposures of the buried ice surface beneath supraglacial debris in Region 2 document an overall pattern of relatively clean ice (0.5–1% debris by volume) alternating with narrow bands of significant englacial debris (each band containing up to 35% debris by volume). The regions of concentrated englacial debris occur precisely where GPR data show inclined reflectors intersecting the glacier surface (e.g., MIR 1–6); clasts in these bands dip up valley, at angles between 7 and 25°, and, on close inspection, appear to rest directly on top of small pebbles; the latter produces an inverse grading pattern that is reminiscent of the

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**Figure 9.** Transverse radar profile \( C_m - C_m' \) on Mullins Glacier, collected with the 200 MHz antenna in distance mode (see Figure 4 for transect location). (top) Unmigrated radar data. (middle) Migrated data with a Hilbert magnitude transform applied. (bottom) Sketch of major radar reflectors and physical interpretation; also shown are the locations of corresponding arcuate surface discontinuities (ASD).
stacking of clasts at the ice surface in Region 1. Inspection of clasts dug out from inclined bands reveals that most display impact scars identical to those observed on the surface of clasts exposed in Region 1. No glacially striated or faceted clasts were observed in the inclined debris bands (or elsewhere within Mullins Glacier). Lastly, exposures of buried ice across Region 2 reveal the presence of near-vertical cracks, each typically less than 2 cm wide and filled with stained quartz sand and dolerite grus (see also section 4.4.5.3 below).

Isotopic analyses of glacier ice (from 0.5 to 1 m depth) recovered from Region 2 fall along a local meteoric water line (LMWL) when plotted on a graph of \(\delta^{18}O\) versus \(\delta^D\) (Gooseff et al., 2006), with a best fit line yielding a slope of \(\delta^D = 8.0 \times \delta^{18}O - 2.14\) \((R^2 = 0.97;\) Figure 10). Deuterium excess values are both slightly negative and slightly positive with the majority clustering around zero \((\text{min} = -5.2\%o, \text{max} = 3.3\%o, \text{mean} = -0.6\%o)\). Analytical precision is ±0.1%. Isotope values are presented as permil (‰) relative to Vienna SMOW. Deuterium excess values were calculated as \(d = \delta^D - 8 \times \delta^{18}O\) after Dansgaard [1964]. Stable isotopic composition, electrical conductivity (EC), and total dissolved solids (TDS) plotted for ice samples from Mullins and Friedman Glaciers. EC values have been adjusted to a reference temperature of 25°C using a standard conversion factor 1.9% °C/°C. EC precision is ±1.0%. Data are also presented as TDS where TDS = EC × 0.51.

Figure 10. Stable isotopic data and electrical conductivity measurements from ice recovered along Mullins and Friedman Glaciers. (a) Plot of \(\delta^D\) versus \(\delta^{18}O\). Also plotted is the local meteoric water line (LMWL) \(\delta^D = 7.75 (\delta^{18}O) - 7.24\) [Gooseff et al., 2006] and the global mean water line (GMWL) is \(\delta^D = 8.0 (\delta^{18}O) + 10\) [Craig, 1961]. (b) Deuterium excess versus \(\delta^D\) for samples plotted in Figure 10a. Deuterium excess values are both slightly negative and slightly positive with the majority clustering around zero \((\text{min} = -5.2\%o, \text{max} = 3.3\%o, \text{mean} = -0.6\%o)\). Analytical precision is ±0.1%. Isotope values are presented as permil (‰) relative to Vienna SMOW. Deuterium excess values were calculated as \(d = \delta^D - 8 \times \delta^{18}O\) after Dansgaard [1964]. (c) Stable isotopic composition, electrical conductivity (EC), and total dissolved solids (TDS) plotted for ice samples from Mullins and Friedman Glaciers. EC values have been adjusted to a reference temperature of 25°C using a standard conversion factor 1.9% °C/°C. EC precision is ±1.0%. Data are also presented as TDS where TDS = EC × 0.51.

4.4.4.3. Geomorphic Data

The thin supraglacial debris of Region 1 increases steadily in average thickness from ~8 to 20 cm (Figure S7b in the supporting information) and is arrayed in arcuate ridges and steps (Figures 3, 4, and 6). We refer to each notable step/ridge by the nongenetic phrase “arcuate surface discontinuity,” or ASD; each ASD is numbered sequentially down valley, with number 1 occurring at the onset of uniform debris cover (Figures 3 and 6).

Region 2 of Mullins Glacier contains six ASD (Mullins ASD 1 through 6). In plan view, Mullins ASD 1–5 all display an elongated, arcuate pattern, with long tails extending several kilometers up valley; these tails all converge toward similar points very near the valley headwall. Mullins ASD 6, which marks the down valley boundary with Region 3, is unique in overall height (10 m relief) and form (lobate front), but it too can be traced well up valley. Each ASD on Mullins Glacier occurs where GPR data show that inclined reflectors intersect the ice surface.

Mullins ASD show the following subtle, but consistent stepwise transitions (Table 3 and Figures 11 and 12):

1. **Slope change**. The buried ice surface, and overlying debris, rise toward each ASD; this forms an asymmetrical step/ridge.
<table>
<thead>
<tr>
<th>Arcuate Surface Discontinuities (ASD)</th>
<th>General Mullins Glacier (Up Valley : Down Valley)</th>
<th>Friedman Glacier (Up Valley : Down Valley)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Surface debris</strong></td>
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<td>Increase</td>
</tr>
<tr>
<td>Thickness (cm)</td>
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<td>1 : 15</td>
</tr>
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<td>Sand (wt %)</td>
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<td>81 : 71</td>
</tr>
<tr>
<td>Gravel (wt %)</td>
<td>45 : 67</td>
<td>34 : 16</td>
</tr>
<tr>
<td>Lg gravel (N)</td>
<td>38 : 30</td>
<td>8 : 10</td>
</tr>
<tr>
<td><strong>Mesoscale features</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polygon flatness (%)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polygon flatness (%)</td>
<td>Increase</td>
<td>Increase</td>
</tr>
<tr>
<td>Vertical relief (cm)</td>
<td>15 : 20</td>
<td>25 : 10</td>
</tr>
<tr>
<td>Diameter (m)</td>
<td>Increase</td>
<td>NA : 22</td>
</tr>
<tr>
<td>(1)</td>
<td>NA : 2.7</td>
<td>5 : 3</td>
</tr>
<tr>
<td>(2)</td>
<td>0.25 : 1.4</td>
<td>0.5 : 2</td>
</tr>
<tr>
<td>Local slope (deg)</td>
<td>Increase</td>
<td>Increase</td>
</tr>
<tr>
<td>(3)</td>
<td>0 : 0 : 0</td>
<td>0 : 0 : 0</td>
</tr>
<tr>
<td>Ridge? Ridge &quot;couplet&quot; formation</td>
<td>Ridge</td>
<td>Ridge</td>
</tr>
<tr>
<td>Ice</td>
<td>Increase</td>
<td>Increase</td>
</tr>
<tr>
<td>Shallow ice debris content (vol %)</td>
<td>0.5 : 1.2</td>
<td>0.5 : 1.2</td>
</tr>
</tbody>
</table>

*No data indicated by -.* Average values are given for the zones immediately up valley of the ASD and down valley of the ASD, separated by a colon. The area that is considered to be immediately up valley and immediately down valley of the ASD varies with each parameter (see below) but ranges from 1 to 40 m. The stepwise changes in surface characteristics across ASD 6, which mark the boundary of region 2 and region 3, are not represented here. Instead, they are explained in detail in the text. The dominant stepwise change observed when comparing the area immediately up valley (UV) of and an ASD to the area down valley (DV) of and ASD. Average calculated using the average till thicknesses for excavations within 100 m laterally from the central flow line and within 40 m (up or down valley) of the ASD. The number of samples used for each thickness average varied from N = 3–10. Average calculated using the grain-size fractions for all excavations within 100 m laterally from the central flow line and within 40 m (up or down valley) of the ASD. The number of samples used for each average varied from N = 1–7. Average calculated using the grain-size fractions for all excavations within 100 m laterally from the central flow line and within 50 m (up or down valley) of the ASD. The number of samples used for each average varied from N = 3–10. Average calculated using the grain-size fractions for all excavations within 100 m laterally from the central flow line and within 50 m (up or down valley) of the ASD. The number of samples used for each average varied from N = 1–7. Average calculated using the grain-size fractions for all excavations within 100 m laterally from the central flow line and within 50 m (up or down valley) of the ASD. The number of samples used for each average varied from N = 3–10. Average calculated using the grain-size fractions for all excavations within 100 m laterally from the central flow line and within 50 m (up or down valley) of the ASD. The number of samples used for each average varied from N = 3–10. Count of clasts >16 mm. Average of values for excavations within 30 m (up or down valley) of the ASD. The number of samples for each average varied from N = 2–7. We define the area analyzed for ASD-related patterned ground analysis as the region bound longitudinally by the ASD and a parallel arc 50 m up or down valley of the ASD. Elevation profiles were extracted from 1.5 m LiDAR [Schenk et al., 2004] DEM along the central flow line in the vicinity of each ASD and smoothed with a five-point moving average. The slope values, derived from the elevation profiles, were smoothed across a 10-point moving average to eliminate the impact of patterned ground. The reported up valley and down valley values are averages for 35 m above and below the ASD, respectively. Ice debris content estimated for a 1 × 1 × 0.5 m³ (x × y × z) ice volume exposed at the base of selected pits. Debris content values include both dispersed fine-grain debris and large grain-size debris.
2. **Debris thickness.** The thickness of supraglacial debris increases by 20–30% at, and immediately down valley from, each ASD.

3. **Bulk grain-size distribution.** The concentration of gravel-sized clasts (<16 mm) and measured cobble-and-boulder fractions increases by a factor of ~1.5–2.0 across ASD.

4. **Patterned ground.** Thermal contraction crack polygons are more ubiquitous and possess larger diameters and deeper troughs immediately down valley of ASD.

5. **Debris content of subjacent ice.** The percentage of observed englacial debris increases from its typical value of ~<1% by volume to as much as 35% by volume in the immediate vicinity of ASD. Importantly, this ice lacks increased silt or amber color that, for example, typically characterizes basal ice in alpine glaciers in the Dry Valleys region [Cuffey et al., 2000a; Mager et al., 2009]; see section 6 below.

Finally, we note that minor arcuate topographic features are also present in Region 2 ("secondary lineations" in Figure 3), but these are small, typically discontinuous and do not display the suite of stepwise transitions in physical properties as observed in Mullins ASD 1–5; furthermore, these secondary lineations are not associated with inclined, englacial reflectors.

**Figure 11.** Sedimentological variation in supraglacial debris on Mullins Glacier as a function of proximity to arcuate surface discontinuities (ASD 1–3) (locations highlighted as shaded bars); grain sizes and sampling procedures for gravel and sand as prescribed in Figure S7 in the supporting information. Smooth lines represent a two-point running average. Both grain size and debris thickness display consistent and abrupt changes across ASD.

**Figure 12.** Diagrammatic illustration showing the typical near-surface geomorphology for ASD. (a) Block diagram highlights changes across a typical ASD (see also Table 3). (b) Generalized cross section (longitudinal profile) showing topographic changes associated with ASD.
4.4.5. Region 2: Friedman Glacier

4.4.5.1. GPR Data

Region 2 (Figure 3) extends from 1.7 to 3.4 km on Friedman Glacier and coincides with the abrupt onset of the complete cover of supraglacial debris. The basal reflector, continuing from Region 1, is diffuse and gradually shoals toward a minimum depth of 360 ns at 3.4 km (Figure 8). Six inclined reflectors (FIR 1 to FIR 6; e.g., Friedman Inclined Reflectors 1, 2, and 3) dominate GPR transects across Region 2 (Figure 8); each inclined reflector intersects the ice surface at a specific arcuate surface discontinuity, ASD, e.g., Friedman ASD 1 through 6 (Figures 3 and 8). Well-defined hyperbolic point reflectors are observed between adjacent FIR in both 80 MHz (Figure 8) and 400 MHz (Figure S2 in the supporting information) transects. Details of the six major inclined reflectors are presented below.

FIR 1 originates from just above the basal reflector and ascends upward to intersect the ice surface at Friedman ASD 1 (~1700–1780 m) (Figures 8 and S8 in the supporting information). The smooth inclination of FIR 1 is interrupted at ~1510 m where it shallows for 15 m, producing a short horizontal section before rising again toward the ice surface (Figure 8). The shallowest portion of FIR 1 (0 to ~20 m depth) splits into two reflectors that intersect the ice surface, at similar sub 25° angles, ~80 m apart (Figure S8). The 400 MHz transect shows that at least the uppermost portion of FIR 1 is composed of individual, closely spaced, hyperbolic reflections (Figure S2).

FIR 2 originates at a depth of ~1350 m near the onset of FIR 1 and rises smoothly to intersect the ice surface at ~1900 m, at a slight topographic ridge, e.g., Friedman ASD 2 (Figures 8 and S8); the angle of intersection with the ice surface is ~27°. At the location of transverse profile Ff1–Ff2, FIR 2 is roughly parabolic in shape and reaches a maximum with of ~300 m where it intersects the buried ice surface (Figure 13). At depth, FIR 2 narrows and displays moderate lateral asymmetry, with its central axis offset slightly to the east of the valley centerline (Figure 13).

FIR 3 originates above the basal reflection at ~1550 m (Figures 8 and S8). It extends down valley at a constant bed-parallel orientation for ~450 m at 700 ns depth before rising sharply toward the ice surface at 2100 m; it intersects the ice surface at 2200 m at ASD 3, with an intersection angle of ~36°. In transverse profile Ff1–Ff3 (Figure 13), FIR 3 is ~450 m wide where it intersects the buried ice surface.

FIR 4 is the least distinct of the major inclined reflectors and originates above the basal reflection at ~2150 m (Figure 8). Thereafter, it rises smoothly to intersect the ice surface at 2350 m, at Friedman ASD 4; the angle of intersection with the ice surface is ~25°.

FIR 5 displays the most complex geometry of any of the major inclined reflectors in Friedman Glacier. The basal portion of this reflector is difficult to differentiate from FIR 3, but it appears to originate above the basal reflector near ~1500 m. It intersects the ice surface nearly 1200 m distant, at 2700 m at the location of Friedman ASD 4; the intersection angle with the ice surface is ~13° (Figure 8). Two distinct bed-parallel portions mark FIR 5: the first extends from ~1500 m to 2400 m and the second from 2450 to 2650 m; the latter corresponds to a distinct gain in reflector thickness. The transverse profile across FIR 4, transect Gf1–Gf2 (Figure S9 in the supporting information), is more complex than that of the other reflectors but shows a roughly parabolic shape that is ~300 m wide at the buried ice surface.

FIR 6 defines the boundary between Region 2 and Region 3 on Friedman Glacier. FIR 6 begins at ~2800 m, remains bed parallel for ~500 m, and ultimately rises to intersect the ice surface at a dipping angle of ~16° at 3400 m (Figure 8).

4.4.5.2. Ice Excavations

Exposures of the buried ice surface observed at the base of 26 soil pits in Region 2 of Friedman Glacier document an overall pattern of relatively clean ice (0–1% debris by volume) alternating with narrow bands of significant englacial debris (each band contains as much as 35% debris by volume). The regions with elevated englacial debris occur at ASD and precisely where GPR data show inclined reflectors intersecting the glacier surface (e.g., FIR 1–6); visible clasts in these debris bands dip up valley, at angles between 7 and 25°, and, on close inspection, appear to rest directly on top of small pebbles. Some of the clasts also show impact scars, identical to those observed on clasts at the ice surface in Region 1, but none show striations or faceting. Lastly, exposures of buried ice across Region 2 reveal the presence of near-vertical cracks, each typically less than 2 cm wide and filled with stained quartz sand and dolerite grus. Isotopic and dissolved salt content analysis of ice samples shows no trend related to proximity with ASD (Figure 10).
4.4.5.3. Geomorphic Data

The thin supraglacial debris of Region 2 shows increasing trends in average thickness (from ~6 to 30 cm) and sand-sized fraction in the down-ice direction (Figure S10 in the supporting information). Friedman ASD displays similar stepwise changes in underlying ice surface slope, topography, and grain-size characteristics (Figure 12) as those mapped on adjacent Mullins Glacier (see section 4.4.4.3 above and Table 3). Each ASD on Friedman Glacier occurs where GPR data show that inclined reflectors appear to intersect the ice surface and where englacial debris increases abruptly to >30% by volume. Additional minor arcuate topographic features are also present in Region 2 (secondary lineations in Figure 3), but these are small, typically discontinuous, and do not display the suite of stepwise transitions in physical properties observed in Friedman ASD 1 through 5; furthermore, they are not associated with the inclined, englacial reflectors.

4.4.6. Region 2: Interpretation

We interpret the set of six inclined reflectors (MIR 1 to 6, FIR 1 to 6) as narrow bands of closely spaced rocky debris, hereafter termed inclined debris layers (IDL, 1–6). This interpretation is based on the high signal amplitude of their radar returns, the negative-positive-negative polarity signature of their radar returns, and the significant volume of englacial rocky debris present at their points of surface intersection (e.g., exposed in surface ice at ASD). The basal reflector in Region 2 of both Mullins and Friedman Glaciers is relatively diffuse and most consistent with scattered debris approximately 10–150 ns "thick" and with the bedrock interface at some unknown depth below. Isotopic analyses of near-surface ice from Region 2, which plot on a local meteoric water line, indicate minimal—if any—melting [Gooseff et al., 2006; Sleewaegen et al., 2003; Souchez and Lorrain, 2006] regardless of proximity to ASD and IDL (see Figure 10).

4.4.7. Region 3: Mullins Glacier

4.4.7.1. GPR Data

Region 3 on Mullins Glacier extends from 4.1 to 5.0 km. The 80 MHz longitudinal radar transect A_m–A_mʹ reveals an increase in high-frequency noise and dispersed reflections. Region 3 displays areas of chaotic, closely spaced reflections with near-vertical dipping angles (Figure 6). Though increasingly diffuse with distance down valley, the basal reflector appears to deepen along the glacier axis, from ~260 ns at 4.1 km to a maximum of ~620 ns at 4.4 km before shoaling up to ~500 ns at 5.0 km (Figure 6).
4.4.7.2. Ice Excavations and Ice Core Data
All cores in Region 3 of Mullins Glacier show clean bubbly ice that displays a subtle increase in englacial gravel and cobble-sized clasts (2–3% by volume) relative to cores from Regions 1 and 2 (Figure 7). Sand veins show a marked increase in concentration, especially as observed at the ice surface, and reach a minimum depth of ~14 m (as observed in ice cores). These sand veins occur as subhorizontal to near-vertical layers, taper with depth, and are filled with stained quartz sands and weathered dolerite grus.

4.4.7.3. Geomorphic Data
Region 3 of Mullins Glacier shows an abrupt increase in the thickness of supraglacial debris (Figure S7b) and a concomitant increase in the maturation of thermal contraction crack polygons (Table 2); the latter includes an increase in average polygon diameter and depth of bounding troughs [e.g., Levy et al., 2006; Marchant et al., 2002]. Isotopic analysis of near-surface ice (0.5–1 m depth) fall on the local meteoric water line (LMWL) (Figure 10).

4.4.8. Region 3: Friedman Glacier
4.4.8.1. GPR Data
Region 3 of Friedman Glacier extends from 3.4 to 4.8 km. The 80 MHz radar (Af–Af’) displays an overall increase in noise and dispersed reflections relative to Regions 1 and 2 (Figure 8). No distinct IDLs are present. The basal reflection is vague and difficult to define, though it appears to remain at a nearly constant depth of ~400 ns.

4.4.8.2. Ice Excavations and Geomorphic Data
Region 3 shows an abrupt thickening in supraglacial debris and an increase in the maturation of thermal contraction crack polygons (Table 2). Ice collected within the top 10–50 cm of buried glacier ice show an increase in the concentration of both sand veins and sporadic gravel-and-cobble-sized debris (collectively up to ~2% by volume).

4.4.9. Region 3: Interpretation
The GPR data suggest that Region 3 of both Mullins and Friedman Glaciers contains mostly clean ice, but with an increasing concentration of dispersed debris relative to that found in Regions 1 and 2. The increase in widespread englacial debris is inferred from (1) spurious reflections within the GPR data, (2) direct observation of debris within shallow ice cores, including sand veins and isolated clasts, and (3) visible englacial debris observed at the base of soil pits and within shallow ice pits.

4.4.10. Region 4: Mullins Glacier
4.4.10.1. GPR Data
Ice in Region 4 of Mullins Glacier (4300 m to 4550 m along Am–Am’) (Figure 6) lacks distinct reflectors and instead displays chaotic signal returns; the 80 MHz signal penetration is less than that of Region 3. A distinct basal reflector is not observed.

4.4.10.2. Ice Excavations and Geomorphic Data
Ice exposed at the base of soil pits in Region 4 shows a large increase in the concentration of gravel-and-cobble-sized clasts. Visible englacial debris reaches ~40–50% by volume, and ice coring is nearly impossible. At present, only one shallow core, 1 m deep, has been extracted; the core contains 35% dispersed sand and gravel, with isolated cobbles containing fresh faces and preserved impact scars (percussion marks [Kowalewski et al., 2011]). Supraglacial debris in Region 4 ranges from ~40 cm to 70 cm thick, typically increasing down valley; the debris is cut by well-developed contraction-crack polygons, each with diameters approaching 20 m (Table 2). As noted above, data from satellite interferometry suggest that horizontal ice velocity in Region 4 approaches zero, essentially <2 mm a⁻¹ and indistinguishable from measurement noise [Rignot et al., 2002].

4.4.11. Region 4: Friedman Glacier
4.4.11.1. GPR Data
GPR data from Region 4 of Friedman Glacier (from 2850 m to 3050 m along A_f–A_f’) show chaotic signal returns; the 80 MHz signal penetration depth is significantly reduced relative to that in Region 3 (Figure 8). A distinct basal reflector is not observed.

4.4.11.2. Geomorphic Data
Region 4 commences immediately below the lobate front of Friedman Glacier. The surface exhibits large diameter polygons (Table 2) (20+ m diameter) that cut into unconsolidated debris at least 70 cm in thickness. At present, it is unclear if glacier ice exists beneath this thick debris cover; if so, it is unreachable.
in hand-dug excavations and not resolvable in our GPR data. According to data from satellite interferometry, surface velocities in this region approach zero and become indistinguishable from measurement noise [Rignot et al., 2002].

4.4.12. Region 4: Interpretation

The closely spaced, high-amplitude reflections in the GPR data for Mullins Glacier are interpreted as debris-laden ice of unknown depth. Individual hyperbolic reflectors cannot be identified within GPR transects, suggesting that individual clasts in Region 4 are not spaced sufficiently far to be resolved separately in the 80 MHz data, e.g., most likely <~40 cm apart.

Our assertion that calls for debris-laden ice (e.g., ice volume > pore space) rather than ice-cemented debris (e.g., ice concentrations ≤ pore space) is supported by visual inspection of the buried ice surface at dozens of localities and shallow ice cores that penetrate this region. Friedman Glacier shows similar GPR data (as that observed in Mullins Glacier), but because buried ice was not physically encountered at the base of soil pits in this region, the relative abundance of ice (ice cement versus excess ice, or even a lack of subsurface ice) remains unknown in Region 4 of Friedman Glacier.

4.5. Determining the Depths of GPR Reflectors

In the absence of a common midpoint profile, we determined a representative average radar travel velocity using measured depths to reflectors from ice-core data (refusal) and confirmed this estimate indirectly through iterative tests of hyperbola collapse during GPR data migration. Borehole MCI-08-01 in Region 2 of Mullins Glacier met refusal at an impenetrable layer at 22.7 m (MIR 4) (Figure 14). Correlating this depth with a strong reflector observed in the colocated 400 MHz GPR profile (Km-Km') yields a travel velocity of 0.168 m ns/C0 (dielectric of 3.18) (Figure 14). This value is slightly higher than the velocity value determined for the headwall area of Mullins Glacier (0.167 m ns/C0 [Shean and Marchant, 2010]) and lower than values determined for a debris-covered glacier on James Ross Island, Antarctica (0.17 m ns/C0 [Fukui et al., 2008]). However, it is within the range of typical radar velocities expected for cold glacial ice [Plewes and Hubbard, 2001, and references therein] and provided excellent hyperbola collapse during iterative tests of radar data migration.

Although this calculated radar velocity is robust at the locations of measured boreholes, it is not necessarily representative of heavily debris-laden ice. In these areas, the travel velocity will likely be slower than our reported velocities, and more consistent with other GPR-travel velocities through debris-rich ice, e.g., 0.12 m ns/C0 [Degenhardt and Giardino, 2003], 0.15 m ns/C0 [Lehmann et al., 1998], 0.14 m ns/C0 [Isaksen et al., 2000], and 0.12 m ns/C0 [Monnier et al., 2008]. The net effect of using a single-average radar travel velocity of 0.168 m ns/C0 will be to slightly overestimate the thickness of englacial debris layers and, as a result, slightly overestimate the depths for rays that travel through these layers. For this reason, we consider our depth-conversion estimates as maximum values and assign a conservative error of ~5% for reported values (equivalent to using an average velocity of 0.160 m ns/C0 or a dielectric of 3.51).

Figure 14. (a) On Mullins Glacier, 400 MHz transect Km-Km' collected across the site of ice core MCI-08-01 (see Figures 4 and S1 for location). (left) Unmigrated data; (right) migrated data with a Hilbert magnitude transform applied. The location of the core is indicated by the grey vertical bars. At the location of the borehole, the GPR indicates a major layer of debris at a two-way travel time of ~270 ns. This debris represents MIR-4/IDL-4. (b) Sketch map of MCI-08-01 showing very little debris except for a few cobbles and gravels at 8 and 16 m and ultimately refusal at 22.7 m. The cobble-rich layer intersected at 22.7 m is almost certainly the same layer imaged at 271 ns. This depth equates to an average radar travel velocity of 0.168 m ns/C0.
In summary, using an average radar travel velocity of $0.168 \text{ m ns}^{-1}$, we arrive at the following: Mullins Glacier reaches a maximum thickness of ~125 m near its headwall at 1.1 km in Region 1; it thins to a depth of ~22 m where its ice surface slope increases near the terminus of Region 2 (3.4 km) and thickens again to ~45 m at 4.6 km (ice thicknesses beyond 4.6 km were not measured in this study, but see Shean and Marchant [2010] for additional data). Friedman Glacier reaches its maximum depth of ~105 m at ~1.25 km in Region 1; thereafter, it thins gradually to ~45 m depth at ~3.6 km in Region 3; if ice continues beyond this point [e.g., Kowalewski et al., 2011], it is unresolvable in GPR data using 80 MHz antennas. Both glaciers display parabolic cross-sectional geometries and rest on a base of inferred rocky debris 3–10 m thick (Regions 2–4) or directly on inferred bedrock (Region 1).

4.6. Ground Penetrating Radar, Ice Cores, and Surface Geomorphology: Interpretation Synthesis

Mullins and Friedman Glaciers share similar surface and internal characteristics that permit subdivision into four distinct regions. Region 1 consists of relatively clean glacier ice, >100 m thick, with scattered englacial debris (<1% by volume). This ice rests on smooth reflectors, most probably bedrock, and is overlain by a discontinuous cover of sporadic cobble-and-gravel-sized clasts. Region 2 also contains relatively clean glacier ice, ~100–25 m thick but additionally includes six, parabolic-shaped, inclined debris layers (IDL) composed of concentrated, rocky debris from 1 to 3 m thick. IDL originate from above the glacier bed and intersect the ice surface at marked, arcuate surface discontinuities (ASD) (Figure S11 in the supporting information). ASD typically occur at longitudinal intervals of ~200–400 m and exhibit long tails that extend back toward ice accumulation areas; isotopic analyses and total dissolved salt content of ice show no systematic variation across individual IDL/ASD. Each ASD shows near-identical changes in ice surface slope, supraglacial debris thickness, patterned ground, and changes in bulk grain-size distribution. Regions 3 and 4 lack well-defined IDL and show stagnant to near-stagnant ice [Rignot et al., 2002] with increasing concentrations of dispersed, englacial debris and a concomitant increase in the thickness of overlying, supraglacial debris.

5. Results: Modeling

5.1. Modeled Glacier Thickness

We utilized a simple analytical steady state flow model to generate expected centerline ice thicknesses for comparison with those derived from GPR. This exercise was carried out in order to (1) provide an independent test of our calculated average GPR travel velocity, (2) confirm our estimates of glacier bed depth — especially in the distal portion of each glacier where the basal reflector is diffuse — and (3) provide initial estimates of the ice hardness parameter for use in future numerical modeling studies. A similar approach was adopted by Shean and Marchant [2010] to estimate ice thickness in upper Mullins Glacier and by Rignot et al. [2002] to estimate first-order ice thickness throughout a large portion of the Quartermain Mountains. Similar to Konrad et al. [1999] and Konrad and Humphrey [2000], we use Glen’s flow law and assume flow without basal slip (due to cold-ice temperatures, see below); we also include changes in overburden pressures arising from the presence of supraglacial debris and incorporate a parabolic shape factor [Hooke, 2005] to approximate the effect of lateral drag by valley sidewalls:

$$h = \frac{v(n+1)}{2A(S_{r} \rho_{g} g \sin \alpha)^{n+1}} + \left( \frac{\rho_{d}}{\rho_{i}} \right) d^{n+1} - \left( \frac{\rho_{d}}{\rho_{i}} \right) d$$

(1)

where $h$ is the depth of the deforming ice body, $v$ is the surface velocity, $n$ is the flow law exponent (3), $\rho_{i}$ is the average bulk density of ice (900 kg m$^{-3}$), $\rho_{d}$ is the average density of the supraglacial debris (1800 kg m$^{-3}$), $g$ is the acceleration of gravity, $d$ is the supraglacial debris thickness, $S_{r}$ is a parabolic shape factor, $\alpha$ is the surface slope, and $A$ is the ice-flow hardness parameter. The supraglacial debris thickness (ranging from 0 to 0.5 m) is derived from a linear fit to measured debris depths (Figure S7a) along the centerline of each glacier and the parabolic shape factor is estimated via interpolating the values from Paterson [1994] using valley geometry derived from GIS measurements and longitudinal and transverse GPR-derived ice thicknesses estimates (see sections 4.4.4.1 and 4.5). The value for the ice hardness parameter, $A$, was initially estimated based on the average measured and modeled ice temperatures (section 5.2) and then iteratively tuned to investigate the goodness of fit to GPR ice thickness measurements.
This simple steady state analytical model is successful in broadly duplicating GPR-derived ice thicknesses along the central flow line of each glacier; results confirm maximum ice thicknesses of ~125 m and ~105 m for the studied portions of Mullins and Friedman Glaciers, respectively (Figure S12 in the supporting information). The best fit for both glaciers occurs with a constant ice hardness flow parameter, $A$, equal to $-7.7 \times 10^{-25}$ s$^{-1}$ Pa$^{-1/3}$. This value of $A$ is slightly higher than what would be expected of cold polar ice [e.g., Paterson, 1994] but is within the range of values calculated for Mullins Glacier ($1.27 \times 10^{-25}$ to $2.34 \times 10^{-24}$ s$^{-1}$ Pa$^{-1/3}$) via seismic depth modeling by Shean et al. [2007] and broadly consistent with the value used for $-23^\circ$C ice in central Beacon Valley ($1.36 \times 10^{-24}$ s$^{-1}$ Pa$^{-1/3}$) by Rignot et al. [2002].

### 5.2. Modeled Thermal Regime

Although the thicknesses and environmental settings of both Mullins and Friedman Glaciers are consistent with cold-based glacial thermal regimes, we explicitly calculate their current maximum basal temperature in order to determine the magnitude of temperature change necessary to reach polythermal basal conditions. Our estimates for the basal thermal regime of both Mullins and Friedman Glaciers are derived from ice thickness estimates (GPR data/modeling) and shallow ice temperatures derived from borehole measurements and modeling. First, we consider a vertical column of ice in one dimension, ignoring strain heating and advection, whereby the heat transfer equation reduces to a simple linear form with a slope of $G/k$, where $G$ is the geothermal flux and $k$ is the average thermal conductivity [Paterson, 1994]. When this linear approximation is fit to borehole temperatures at 10 m and 30 m at MCI-09-03, we arrive at an estimated geothermal flux of 105 mW/m² and a maximum estimated basal ice temperature of $-17.4^\circ$C at 125 m (the measured maximum depth of Mullins Glacier) (Figures S5 and S13 in the supporting information). Employing the full range of reported values of $G$ for the MDV region ($68$–$105$ mW/m²), we derive an envelope for basal ice temperatures at 125 m depth of between $-17$ and $-19^\circ$C. We also modeled the thermal regime using the thermal solution within the thermomechanically coupled Ice Sheet Systems Model [Larour et al., 2012] in order to investigate any potential heat contributions from strain heating/advection. As expected for these thin and slow-moving glaciers, we found contributions from these sources to be negligible (Figure S13). The only model scenarios that produce basal ice temperatures near the pressure melting point are those that require mean annual ice surface temperatures well above modern values (e.g., an increase in mean annual atmospheric temperature by $-17^\circ$C) and unrealistically high values of $G$ (approaching $-180$ mW/m²).

### 5.3. Modeled Rockfall Runout Distance

In an effort to understand better the potential sources for englacial and supraglacial debris on Mullins and Friedman Glaciers, we utilized the numerical simulations within CRSP-3d (Colorado Rockfall Simulation Program) [Andrew et al., 2012; Harany et al., 2013] to model run out distances for falling clasts sourced along the headwall of Mullins Glacier. At the core of CRSP-3d is a discrete element method (DEM) simulation of rock velocity (translational and rotation in three dimensions) and contact forces between rock and slope [Andrew et al., 2012; Bartingale et al., 2009; Leine et al., 2013]. Falling rocks are defined by shape, size, and density, and slope material is parameterized by physical surface roughness and a hardness coefficient—a value that subsumes both the coefficient of restitution and a dampering coefficient; see additional model parameterizations and run scenarios in Figure 15.

The modeled rockfall run out distances using the CRSP-3d model are provided in Figure 15. As expected, rockfall from isolated rock outcrops yields narrow, linear tracks of concentrated debris, rather than widespread layers across the glacier surface. In addition, the modern headwall geometry and ice slope characteristics favor long run out distances, with deposition primarily within the lower accumulation area (~70%) and the upper ablation area (30%). Debris trapped in the accumulation area generally includes the smaller clast sizes representative of most Mullins supraglacial debris (<2.0 m). This debris would become buried beneath subsequent snow and ice to undergo rotation and transportation along internal flow lines [Berthling et al., 2000]; those rocks that initially come to rest out on the ablation area include the largest clast sizes (>2.0 m); these clasts would travel solely as supraglacial debris (Figures 16).

### 6. Discussion

We evaluate potential origins for englacial and supraglacial debris within Mullins and Friedman Glaciers. We begin with supraglacial debris, for which a candidate hypothesis for its origin has already been proposed...
Kowalewski et al., 2011, and references therein]; we then evaluate potential origins for the enigmatic layers of inclined rocky debris (IDL). We end with an assessment of whether cold-based, debris-covered glaciers containing distinct englacial debris bands can yield reliable records for long-term environmental change.

6.1. The Origin of Supraglacial Debris on Mullins and Freidman Glaciers

The prevailing hypothesis for the origin of supraglacial debris on Mullins and Friedman Glaciers is rockfall sourced from rock cliffs at valley heads [Kowalewski et al., 2011; Rignot et al., 2002; Shean and Marchant, 2010] (Figure 16). This model is entirely consistent with data presented above. However, the relative percentage of rockfall that lands directly onto ice ablation areas (and travels solely as supraglacial debris) versus the percentage that lands in ice accumulation areas—and thereby includes an additional component of englacial transport [Kowalewski et al., 2011; Marchant et al., 2013]—is currently unknown. The distinction is important, not only for understanding transport pathways but also for assessing potential origins for the inclined layers of concentrated englacial debris bands (IDL). We end with an assessment of whether cold-based, debris-covered glaciers containing distinct englacial debris bands can yield reliable records for long-term environmental change.

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Based on results from GPR data, the measured thickness of supraglacial debris, and the concentration of debris observed in shallow ice cores, it appears that most of the supraglacial debris on Mullins and Friedman Glaciers could have formed directly from sublimation of ice containing rocky debris. For example, if we (1) accept 1% as the maximum concentration of dispersed englacial debris (by volume) in Regions 1 and 2 of Mullins and Friedman Glaciers (excluding IDL, see below), (2) assume that ice sublimation decreases from $-10^{-1}$ mm a$^{-1}$ to $-10^{-2}$ mm a$^{-1}$ across Region 2 [Kowalewski et al., 2011]; that horizontal ice flow for Region 2 is $\leq 40$ mm a$^{-1}$ [Rignot et al., 2002]; and (3) consider surface erosion is essentially negligible over the time scales considered here [Morgan et al., 2010a, 2010b], then nearly 70% of the observed thickness of...
supraglacial debris in Region 2 can be attributed to sublimation of underlying ice with 1% scattered englacial debris (by volume). If correct, then the remaining 30% of the supraglacial debris would likely come from rockfall deposited directly on top of ice ablation areas and from the addition of concentrated englacial debris located within inclined debris layers (IDL, see below).

To test this hypothesis, we turn to the results of the modeled expected rockfall runout distances (section 5.3), which show, to a first order, that given the present-day ice surface slopes and headwall geometries, the majority of rockfall derived from exposed rock outcrops at valley heads would come to rest on ice accumulation areas and subsequently follow englacial transport pathways; only exceptionally large clasts, \( \geq 2-3 \, \text{m in diameter (rare on Mullins and Friedman Glaciers),} \) would typically be deposited in ice ablation areas (Figure 15). The subtle increase in the thickness of supraglacial debris across lobate ridges and steps on Mullins and Friedman Glaciers (a 10–20% increase across arcuate surface discontinuities, ASD), as well as the consistent change in grain size across these boundaries, suggests that sublimation of ice in IDL adds an important but relatively minor contribution to supraglacial debris. Taken together, the results suggest that the vast majority of supraglacial debris (~70% of its thickness) is derived from sublimation of underlying ice with scattered englacial debris, with the remaining ~30% derived from a combination of rockfall events that result in some component of long-distance runout and, locally, by the addition of rocky debris within IDL.

6.2. The Origin of Inclined Debris Layers (IDL) Within Mullins and Friedman Glaciers

Inclined layers of englacial debris are relatively uncommon features in alpine glaciers but have been observed consistently in debris-covered glaciers and traditional rock glaciers [Berthling et al., 2000; Degenhardt, 2009; Degenhardt and Giardino, 2003; Degenhardt et al., 2003; Isaksen et al., 2000; Monnier et al., 2008]. Although purported origins vary, most posited origins require some type of basal entrainment mechanism followed by active transport toward the ice surface. Two related models applicable to Mullins and Friedman Glaciers are shearing and thrusting of subglacial ice- rich permafrost [e.g., Chinn and Dillon, 1987; Fukui et al., 2008; Strelin and Sone, 1998] and basal entrainment beneath cold-based ice [e.g., Atkins, 2013; Atkins et al., 2002; Cuffey et al., 2000a; Mager et al., 2009]. A third model, which involves the concentration of rockfall debris in ice accumulation areas, followed by subsequent burial beneath snow and ice, and englacial
transport, is also possible [e.g., Berthling et al., 2000; Degenhardt, 2009; Degenhardt and Giardino, 2003; Degenhardt et al., 2003; Isaksen et al., 2000; Monnier et al., 2008]. The merits of all three models as potential origins for IDL in Mullins and Friedman valleys are presented below.

6.2.1. Origin of IDL by Basal Shearing and Thrusting of Preexisting, Ice-Rich Debris

Although shearing and thrusting of subglacial debris has been postulated to explain the origin of debris layers in polythermal debris-covered glaciers [e.g., Chinn and Dillon, 1987; Fukui et al., 2008; Strlin and Sone, 1998], it has yet to be applied to cold-based alpine glaciers like Mullins and Friedman Glaciers (e.g., with basal temperatures <16°C). The basic assumption of this model is that heterogeneous ice movement, especially across areas of enhanced/reduced basal flow typically associated with thermal transitions, may induce shear stresses and thrusts [Nye, 1952] that could potentially provide both an entrainment mechanism and transport vector for uplifting subglacial debris [Alley et al., 1997; Chinn and Dillon, 1987; Fukui et al., 2008; Glasser et al., 2003; Hambrey et al., 1999; Hubbard et al., 2004]. In addition to differential basal flow, the model requires the persistent availability of ice-rich subglacial debris. For Mullins and Friedman Glaciers, the model must also account for punctuated entrainment that is uniform in both space and time across two separate glaciers. Our GPR data suggest that although ice-rich debris may underlie glacier ice in Regions 2–4, such debris is most probably absent from Region 1. Rather, GPR data from Region 1—where IDL originate—show profiles with
smooth, basal returns, most plausibly interpreted as coherent bedrock. Where these smooth reflectors are most visible in GPR profiles (Figures 6 and 9), they appear to lie well below the onset of mapped IDL, suggesting individual IDL are most probably sourced above the bed. Additionally, given the inherent complexities in bedrock topography, ice thicknesses, and ice surface slopes, it is unlikely that punctuated basal shearing would be consistent across two separate glaciers. One possibility is that past episodes of climate warming may have temporarily yielded polythermal ice conditions, leading to episodic basal entrainment and thrust formation [e.g., Hubbard et al., 2004]. However, there is little evidence for recent climate warming of the magnitude to produce polythermal conditions (~17°C increase for the current ice thickness; Figure S13); rather, local evidence calls for enduring hyperarid, cold-desert conditions for the last several million years and extreme landscape stability over this duration [Marchant et al., 2002, 2013; Schaefer et al., 2000; Sugden et al., 1995b]. In support of this assertion, we highlight the observation that clasts within IDL do not exhibit striations or faceting that would be consistent with basal entrainment. For these reasons, we consider basal shearing and thrusting of preexisting, ice-rich debris to be an unlikely origin for IDL within Mullins and Friedman Glaciers.

6.2.2. Origin of IDL Through Basal Entrainment Beneath Cold-Based Ice and Subsequent Transport Along Shear Planes

Recent studies have shown that minor erosion is possible beneath cold-based ice [e.g., Atkins, 2013; Atkins et al., 2002; Cuffey et al., 2000a]. Atkins et al. [2002] and Atkins [2013] have demonstrated convincingly that, in places, cold-based ice can rotate and transport loose debris across bedrock, producing gouges and scratches and depositing thin gravelly till. Others have postulated far more significant erosion and basal entrainment; however, on close inspection, erosion and entrainment in these latter cases has been localized at the margins of thawed patches in polythermal ice, rather than the broad areas of cold-based ice [e.g., Chinn and Dillon, 1987]. Finally, Cuffey et al. [2000a] modeled erosion rates beneath a fully cold-based alpine glacier in the MDV (Merville Glacier, with basal temperatures everywhere ≤ −17°C) and calculated rates on the order of 10−7 m Ma−1.

In further consideration of a basal-entrainment mechanism for IDL within Mullins and Friedman Glaciers, we first discuss whether the requisite erosion rates and subsequent entrainment of rocky debris in IDL would greatly exceed postulated and calculated rates of erosion beneath typical cold-based ice [Cuffey et al., 2000a]. Assuming a conservative ice flow rate of 35 mm a−1 [Rignot et al., 2002], an average IDL thickness of 2 m, and an average debris concentration in IDL of ~25% by volume, the volume flux of IDL exiting Region 1 is on the order of 0.0175 m3 m−2 s−1. Averaged across the entire bed of Region 1 (required by the widespread distribution of IDL observed in transverse GPR profiles), this flux would require a glacier-wide erosion rate on the order of 4 × 10−5 m/yr, roughly 2 orders of magnitude greater than that reported for the cold-based Merville Glacier [e.g., Cuffey et al., 2000a]. Furthermore, because entrainment would likely be localized and not uniform across all of Region 1, this calculation represents a conservative minimum erosion rate and the localized entrainment rate would be much higher.

In addition, we note that the well-documented, tell-tale signs for basal entrainment beneath cold-based ice are consistently lacking from IDL in Mullins and Friedman Glaciers. Missing indicators for debris entrainment beneath cold-based ice include the following: the presence of ubiquitous amber-colored ice [Cuffey et al., 2000a; Mager et al., 2009; Samyn et al., 2005; Slewaegen et al., 2003; Tison et al., 1993]; a several fold increase in the concentration of total dissolved solids [Cuffey et al., 2000b; Holdsworth, 1974; Holdsworth and Bull, 1970; Mager et al., 2009]; an increase in silt and dispersed fine debris (as much as 5%) [Fitzsimons, 2003; Mager et al., 2009]; striated or glacially molded clasts; or the appearance of “muddy clots” within otherwise fine-grained debris [Tison et al., 1993]. None of these characteristics are present within IDL of Mullins and Friedman Glaciers. In fact, apart from the noted increase in the abundance of gravel-and-cobble-sized clasts in mapped IDL’s, the associated ice is identical in isotopic composition, bulk-dissolved solute content, and visible appearance to that observed elsewhere throughout Regions 1–4 of Mullins and Friedman Glaciers.

Finally, and perhaps most significantly, we note that typical cold-based alpine (nondebris-covered) glaciers in the MDV lack the englacial debris bands that so characterize the surface and internal structure of Mullins and Friedman Glaciers. If cold-based ice were capable of significant basal entrainment followed by subsequent upward movement of debris along thrust planes [Alley et al., 1997; Fukui et al., 2008; Glasser et al., 2003; Hambrey et al., 1999], then we would expect to find similar IDL within other cold-based alpine glaciers throughout the Dry Valleys, which is not the case [Chinn, 1980, 1985; Fountain et al., 2006]. Some englacial debris bands do occur in glaciers of the MDV, but these bands are invariably found near glacier snouts [Chinn,
Friedman Glaciers. The et al. 2000; multiunit structures within debris-covered glaciers and rock glaciers outside of Antarctica [Fi

Other potential basal entrainment mechanisms for IDL are equally problematic. These might include the trapping of debris within basal crevasses [Bennett et al., 1996; Ensminger et al., 2001], the folding of debris-laden basal ice [e.g., Glasser et al., 1998], and, as noted above, the overriding of terminal debris aprons [Hooke, 1973; Shaw, 1977]. However, these processes are typically only active within thin terminal regions of polythermal glaciers and therefore are not likely applicable to Mullins and Friedman Glaciers.

Notwithstanding the above, we wish to emphasize that the data presented do not preclude the possibility of localized shear along established IDL; however, they are first formed. In fact, the discrete rheological contrasts presented by debris-laden IDL [Arenson et al., 2002; Colaprete and Jakosky, 1998] in otherwise clean glacier ice would provide preferred planes for differential shear under longitudinal compressive stress [Hooke et al., 1972]. Thus, shear and enhanced vertical transport may occur along some of the IDLs in Mullins and Friedman Glaciers [e.g., Shean and Marchant, 2010], but this does not imply that they formed by this mechanism. Minor differential shear would provide one explanation for very subtle changes in horizontal ice surface velocity observed in InSAR data [Rignot et al., 2002]. Future analysis of ice fabrics may help illuminate the nature and location of any potential shear [e.g., Alley, 1988] associated with IDL.

6.2.3. Rockfall Origin for IDL Within Mullins and Friedman Glaciers

As noted above, the vast majority of alpine glaciers in the MDV lacks supraglacial debris; the few debris-covered glaciers that do exist form only in the lee of cliffs, principally below cliffs of Ferrar Dolerite and Beacon Supergroup sandstones [Marchant et al., 2013]. Moreover, all clean-ice alpine glaciers (e.g., those without supraglacial debris) in the MDV appear to lack IDL [Chinn, 1980, 1985; Fountain et al., 2006; Mager et al., 2009, and references therein]. Taken together, these simple observations suggest that the origin of supraglacial debris and IDLs may be related and both may require rockfall from valley headwalls.

In consideration of this model, we note that changes in rockfall—and subsequent burial of debris in ice accumulation areas—has often been attributed as a causal mechanism for internal debris layers and multiunit structures within debris-covered glaciers and rock glaciers outside of Antarctica [Berthling et al., 2000; Degenhardt, 2009; Degenhardt and Giardino, 2003; Degenhardt et al., 2003; Isaksen et al., 2000; Monnier et al., 2008]. We postulate two end-member scenarios for a rockfall-centric origin for IDL in Mullins and Friedman Glaciers. The first involves high-magnitude rockfall events that would place significant debris on ice accumulation areas; such an event, or series of events closely spaced in time (e.g., <~10 years apart) would yield a geologically instantaneous rocky layer that could ultimately be buried beneath subsequent snow and ice, producing inclined layers of englacial debris [e.g., Berthling et al., 2000] (Figure 16b). We refer to this model as the large-magnitude rockfall model (LMR). In contrast, the second scenario relies on maintaining low rates of rockfall (similar to today’s) but instead requires marked changes in ice accumulation and ablation. A decrease in net ice accumulation would favor lowering of exposed glacier ice near valley headwalls (e.g., ice not protected beneath debris) but negligible lowering for glacier ice buried beneath supraglacial debris. This differential ablation would produce spoon-shaped hollows near valley headwalls [Ackert, 1998; Humlum, 1996; Johnson and Lacasse, 1988; Potter, 1972; Whalley, 1979] that—owing to the continued sublimation of ice with scattered englacial debris and persistent low-amplitude rockfall events—would ultimately become covered with a uniform and widespread rocky lag. The overall grain size, sorting, and spatial arrangement of clasts within IDL, where observed directly in hand-dug ice trenches, tends to mimic the distribution and stacking of large cobbles on top of small pebbles observed in the upper ablation areas of Mullins and Freidman glaciers. This similarity hints at the possibility that IDL may have developed in ice accumulation areas through slow sublimation processes similar to those currently operating in the ablation areas today. Renewed net-ice accumulation would bury these lags, initiate englacial transport, and ultimately produce englacial debris bands (IDL). We refer to this model as the climate-hollow-lag model (CHL).
In assessing the LMR model, we note that the strong correlation between the number and spacing of IDL (and associated ASD) across Mullins and Friedman Glaciers implies regional forcing, rather than the linkage of stochastic rockfall at the heads of Mullins and Friedman valleys. On the other hand, simultaneous detachment of large volumes of rockfall could occur if forced by regional seismic activity [e.g., Reznichenko et al., 2011]. If seismicity was the driving force for IDL, then one would expect to see widespread evidence for recent and catastrophic rockfall beneath all steep cliffs throughout the western Dry Valleys region, which is not the case [Denton et al., 1993; Marchant et al., 1996; Putkonen et al., 2012; Sugden et al., 1995a; Swanger and Marchant, 2007]. Instead, current models for landscape evolution suggest incredible landscape stability, with extremely low rates of sporadic rockfall; the region possesses some of the lowest rates of downslope movement on Earth [Marchant et al., 2013; Putkonen et al., 2012] and minimal bedrock erosion, the latter averaging $<10$ cm/Ma [Summerfield et al., 1999a, 1999b]. Even if synchronous, large-magnitude rockfall events were possible; modeling studies from Okura et al. [2000] suggest that for bedrock geometries comparable to those of Mullins and Friedman valleys, large-magnitude rockfall events would result in long-distance transport; in this case, clasts would likely travel well out onto ice ablation areas and essentially bypass ice accumulation areas, thereby preventing development of IDL (see also Figure 16b). Finally, the uniform thickness and widespread areal extent of IDL, especially as observed in transverse GPR profiles, is difficult to reconcile via large-magnitude rockfall. Deposits from large rockfall events are typically elongated in plan view, with high length-to-width ratios, and are inconsistent with the observed distribution (across the entire glacier width) and uniform thickness of mapped IDL.

One possibility is that regional climate change may induce marked changes in ice accumulation and ablation as well as an increase in the overall frequency rockfall events [Gruber and Haeberli, 2007; Gruber et al., 2004]. Together, these changes could produce widespread lags in upper ablation areas that, with renewed ice accumulation could form IDL. Linkages between climate change and rockfall frequency have been noted for cliffs in subpolar and temperate regions, where changes in the rate of meltwater development have been implicated with climate warming [e.g., Gruber and Haeberli, 2007; Gruber et al., 2004; Huggel et al., 2012; Ravanel and Deline, 2011]. However, such linkages have not been documented for cliffs in the Transantarctic Mountains, where mean annual temperatures are well below zero Celsius ($< -23°C$ in this case) and where the necessity of frequent rockfall events appear at odds with geomorphic data that call for long-term landscape stability throughout the region [Marchant et al., 2013; Putkonen et al., 2012]. On the basis of the above data, we cannot rule out a combination of increased rockfall rate and decreased ice accumulation as co-drivers in the development of IDL, but the most parsimonious explanation is that rockfall rates remain low and relatively constant—as suggested by allied geomorphic data—and that lags form predominantly by reduced ice accumulation and the development of spoon-shaped hollows.

### 6.3. The Potential for Debris-Covered Glaciers in the MDV to Register Long-Term Climate Change

Figure 17 presents a diagraphic illustration of our preferred model for the development of IDL, the Climate-Hollow-Lab model (CHL). As shown in Figure 17b, during episodes of reduced ice accumulation ($t = 2$), ablation is most pronounced at the surface of exposed (unprotected) ice in the upper portions of the glacier [Ackert, 1998; Konrad et al., 1999; Kowaleski et al., 2011; Potter et al., 1998; Scherler et al., 2011]. Such differential ablation forms the localized topographic depression (termed “ablation hollows”) near glacier heads; most of the protected portion of the glacier is little affected and continues to slowly flow down valley. Due to reversed ice surface slopes in the vicinity of the down valley portion of the hollow (Figure 17b), localized ice flow in this region may flow very slowly up valley, partially infilling and elevating the nescient hollow. However, given the low ice temperatures, thin ice conditions, and relatively minor changes in ice surface elevations, any reverse flow into the hollows would be minimal compared to ice loss from sublimation. As the ablation hollow forms, its thickening mantle of surface debris (the evolving lag from sublimation of ice with rocky debris and continued, minor rockfall) forms an insulating layer that halts runaway ice loss. Following a climate transition favoring net ice accumulation ($t = 3$, Figure 17b), the ablation hollow and overlying surface debris are buried. This now-buried debris (IDL) is subsequently deformed and advected down valley [e.g., Berthling et al., 2000; Degenhardt, 2009; Degenhardt and Giardino, 2003; Goodsell et al., 2005; Hooke and Hudleston, 1978; Isaksen et al., 2000; Vaughan et al., 1999]. The processes repeats, forming IDL at similar times in both Mullins and Friedman Glaciers ($t = 4$).
Although the duration and magnitude of climate forcing required to produce discrete IDL is unknown, the extremely slow horizontal ice-flow velocities recorded in InSAR \cite{Rignot et al., 2002}, as well as the reported low sublimation and high potential for long-term ice preservation \cite{Kowalewski et al., 2011}, suggest causation by relatively low-frequency climate perturbations, perhaps on the order of \(10^4\) years.

Our preferred model for the development of IDL builds upon aspects of previous models that invoked some combination of climate change and partially overriding flow lobes in talus-derived rock glaciers \cite[Degenhardt, 2009; Monnier et al., 2008]{}, variations in debris input or ice accumulation/ablation \cite[Barsch, 1996; Olyphant, 1987]{}, and alternating rockfall regimes driven by fluctuating headwall geometries \cite[Ackert, 1998]{}. In our departure from these previous models, we conclude that the salient features of Mullins and Friedman Glaciers that are necessary (if not required) to form IDL (and associated ASD) via the CHL model mechanism include the following: (1) persistent cold-based ice conditions, (2) rockfall source areas in the headwall region, (3) increasingly thick supraglacial debris in the down valley direction, and (4) low overall ablation rates. Therefore, other debris-covered glaciers that meet these criteria may also contain records of long-term climate change. However, we caution that not all surface lineations/ridges on Mullins and Friedman glaciers (or similar debris-covered glaciers elsewhere) can be considered ASD, as defined here, or must be associated with climate perturbations. These locally discontinuous surface ridges that lack associated IDL's and do not show the stepwise changes in surface geomorphology that define ASD \cite[e.g., the secondary lineations shown in Figure 3]{} are most likely related to compressive flow \cite[e.g., Burger et al., 1999; Kääb and Weber, 2004]{}, perhaps exacerbated by changes in flow velocity related to the alternate development of ablation hollows near valley heads, followed by renewed ice accumulation and enhanced longitudinal flow.

7. Conclusions

We used ground-penetrating radar, ice core analyses, geomorphic mapping, and numerical modeling to investigate the surface and internal structure of two cold-based, debris-covered glaciers situated within neighboring valleys in the cold-and-dry Quartermain Mountains of southern Victoria Land, Antarctica. Friedman Glacier, \(\sim100\) m thick at its deepest point near the valley headwall, thins steadily along its length before terminating at 3.8 km. Mullins Glacier thins from \(\sim125\) m thick near its headwall toward a local minimum of \(\sim22\) m at 3.4 km distance; the thickness of Mullins Glacier ice at the termination of GPR coverage (5.2 km down valley) could not be determined due to incomplete penetration of the radar signal.

A simple application of a modified Glen’s flow law reasonably predicted measured ice thicknesses and flow velocities calculated from InSAR \cite{Rignot et al., 2002}{.} Both Mullins and Friedman Glaciers are cold based and frozen to their beds.

Both glaciers are overlain by a layer of dry supraglacial debris (8–75 cm thick), each of which is marked with near-identical patterns of arcuate surface ridges and steps (ASD); each ASD, in turn, displays uniform changes in bulk-sediment grain size, meter-scale surface topography, and the geometry and maturity of surface polygons.

Results from 24 km of ground-penetrating radar (80, 200, and 400 MHz antennas) show that each ASD is underlain by a distinct layer of rocky englacial debris, inclined up glacier, and containing as much as 35% debris by volume. Between these inclined debris layers, ice in Mullins and Friedman Glaciers is nearly devoid of englacial debris \(<1\% by volume); the most likely source for the scattered englacial debris \(<1\% by volume) is sporadic rockfall at valley headwalls.

The similarity in number and pattern of englacial debris bands and corresponding ASD across both glaciers, along with model results and observations that call for negligible basal entrainment as a source for englacial debris, is best explained by episodic environmental change at valley headwalls. We envision three distinct components to this model: (1) continued low-amplitude rockfall (similar to today’s rate) to produce scattered englacial debris; (2) a reduction in net ice accumulation that favors a marked reduction in the ice surface elevation of (unprotected) ice in former ice accumulation areas but negligible ice surface lowering beneath debris-covered (protected) portions of each glacier; ultimately, this differential ablation produces a spoon-shaped hollow near valley headwalls that, eventually, is capped with rocky lags produced from continued ice sublimation and low-amplitude rockfall events; (3) renewed net ice accumulation buries these lags, initiates englacial transport, and ultimately produces the observed englacial debris bands and accompanying arcuate surface discontinuities.
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If the model is correct, the implication is that cold-based, debris-covered glaciers with distinct englacial debris bands and corresponding changes in surface morphology may record environmental change. Given the propensity for long-lived debris-covered glaciers in Antarctica, this opens up the possibility that a subset of debris-covered glaciers could serve as reliable proxies for long-term climate change, potentially covering the last several hundred thousand years duration.


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