Entrainment at cold glacier beds

K. M. Cuffey Department of Geography, 501 McCone Hall, University of California, Berkeley, California 94720-4740, USA
H. Conway, A. M. Gades, B. Hallet — Geophysics Program and Department of Geological Sciences, Box 351650, University of Washington, Seattle, Washington 98195, USA
R. Lorrain Departement des Sciences de la Terre et de l'Environnement, Université Libre de Bruxelles, B-1050 Bruxelles, Belgium
J. P. Severinghaus Scripps Institute of Oceanography, 9500 Gilman Drive, La Jolla, California 92093, USA
E. J. Steig*, B. Vaughn, J. W. C. White — Institute of Arctic and Alpine Research, University of Colorado, CB450, Boulder, Colorado 80309, USA

ABSTRACT

Here we present measurements of the gas content and isotopic composition of debris-rich basal layers of a polar glacier, Meserve Glacier, Antarctica, which has a basal temperature of –17 °C. These measurements show that debris entrainment has occurred without alteration of the glacial ice, and provide the most direct evidence to date that active entrainment occurs at the beds of cold glaciers, without bulk freezing of water. Entrainment at subfreezing temperatures may have formed the U-shaped trough containing Meserve Glacier. In addition to possibly allowing some cold-based glaciers to be important geomorphic agents, entrainment at subfreezing temperatures provides a general mechanism for formation of the dirty basal layers of polar glaciers and ice sheets, which are rheologically distinct and can limit the time span of ice-core analyses. Furthermore, accumulating evidence suggests that geomorphologists should abandon the assumption that cold-based glaciers do not slide and abrade their beds.

Keywords: glacial erosion, glacial entrainment, glacial valleys, ice cores, Antarctic Dry Valleys.

INTRODUCTION

Dirty basal ice layers are present at many locations beneath polar glaciers and ice sheets having basal temperatures that are lower than bulk pressure-melting temperature (Holdsworth, 1974; Koerner, 1989; Fitzsimons, 1996; Gow et al., 1997). These layers are important because they usually deform more rapidly than clean ice and can contribute significantly to the flow of the ice mass (Echelmeyer and Zhongxiang, 1987; Dahl-Jensen and Gundestrup, 1987); in addition they may limit the time span of useful paleoclimate information inferred from ice-core analyses. In some cases, these layers clearly are relicts of past climatic conditions and ice-sheet configurations that are drastically different from the modern ones, as under central Greenland (Souchez, 1997; Souchez et al., 1995). In general, however, dirty basal layers will be present irrespective of such historical circumstance only if entrainment of debris into basal ice occurs actively at subfreezing temperatures.

In this article we present a case study (a continuation of work begun by Holdsworth and Bull, 1970) that provides the best example to date, as far as we know, of entrainment occurring at subfreezing temperatures. In addition, this case study provides an opportunity to challenge the assumption that cold-based glaciers do not accomplish significant geomorphic work, and hence that landforms of glacial erosion beneath subtemperate ice masses are relicts of past warm-bed conditions (e.g., Richardson and Holmlund, 1996). Furthermore, we discuss entrainment mechanisms and emphasize that recent evidence from both physics and glaciology implies an active role for liquid water-dependent processes in subfreezing ice. In this context, motivated in particular by our recent high-resolution continuous measurements of slow sliding at the Meserve Glacier base (Cuffey et al., 1999), we renew the challenge (Shreve, 1984) against the common assumption that cold-based glaciers do not slide or abrade their beds (e.g., Sugden and John, 1976; Summerfield, 1991; Denton et al., 1993; Gow et al., 1997), and hence that striated rock surfaces are evidence for past warm-bed conditions. Thus, although we do not challenge the idea that cold-based ice masses are weak erosive agents relative to temperate ones, we argue that the interaction between cold-based glaciers and their beds may be more important than commonly assumed in a variety of glaciologic and geomorphic problems.

To discuss these problems, the following terminology distinction is used. The term subfreezing means at temperatures below the bulk pressure-melting point. Near an interface, liquid water can exist in stable equilibrium at subfreezing temperatures. Thus the proper physical distinction (Shreve, 1984) is not melting versus frozen, but rather net melt versus no net melt. At subfreezing conditions there is no net melt, the volume of water being fixed by physics and geometry of interfaces.

EVIDENCE FOR ENTRAINMENT AT SUBFREEZING TEMPERATURES?

Mechanisms for entrainment of coarse rocks at subfreezing glacier soles have been proposed (Sugden and John, 1976), but direct evidence of subfreezing entrainment is lacking. Observations of rock debris entrained in the terminal ice of polar glaciers (e.g., Mercer, 1971) do not provide the necessary evidence because the mechanism for entrainment in these cases is ambiguous. Entrainment could occur by melt and refreeze at the margins of these glaciers prior to advance, by active shearing of overridden stony permafrost, from supraglacial sources, or by normal melt-freeze entrainment processes in temperate ice (Alley et al., 1997), either far upflow or at some time in the past when the ice was thicker or the climate warmer.

Analyses of the composition of dirty basal layers beneath polar glaciers have mostly concluded that entrainment occurred under warm-bed conditions via bulk melting and refreezing, whether current (Sugden et al., 1987; Zdanowicz et al., 1996; Souchez et al., 1988; Lorrain et al., 1999), or past (Souchez et al., 1995; Koerner, 1989). An exception is the work of Tison et al. (1993), who suggested that debris travels from the bed upward along shear bands at subfreezing temperatures. This suggestion is not well supported because their gas and isotopic data reveal significant alteration by water.

CASE STUDY: MESERVE GLACIER

Entrainment

In the austral summer of 1995–1996, we excavated a 20-m-long tunnel through basal ice of Meserve Glacier, an alpine glacier in Wright Valley, Antarctica. Earlier studies here (Holdsworth and Bull, 1970) revealed dirty basal layers despite the low temperature of –17 °C, and we wished to study these using gas and isotopic composition techniques not available to Holdsworth and colleagues. The dirty basal layers contain dispersed fine rock particles (to 2.5% silt and fine sand) and larger boulders, and have a yellow-brown amber hue (Munsell designation: 2.5Y 4/4 to 7/4).

The margin of Meserve Glacier's ablation zone is a 16-m-high vertical cliff (see Holdsworth, 1974), from which ice blocks fall to accu-

*Present address: Department of Earth and Environmental Sciences, University of Pennsylvania, 251 Hayden Hall, 240 South 33rd Street, Philadelphia, Pennsylvania 19104, USA.

mulate in an apron of marginal debris at the cliff foot. This apron also contains windblown snow and refrozen meltwater. On the warmest days of summer, the glacier surface and this marginal apron melt. Recent advance of the glacier and its amber basal layers over this marginal debris has added a third, basal stratigraphic layer within 20 m of the margin (Holdsworth and Bull, 1970; Fig. 1). The boundaries between these units are blurred by physical mixing, manifested as interfingering and recumbent folding of distinctive layers. The amber layers are in direct contact with the bed except in this 20-m-wide zone near the ice cliff (Holdsworth and Bull, 1970).

The amber layers constitute a clearly distinct facies, which cannot result from mixing of the overlying and underlying ices, the former having 100 times the rock content of the others and a stable isotopic composition 40‰ lighter (in δD) than the subjacent ice but identical to the suprajacent ice.

Gas Composition. As snow becomes denser to form glacier ice on polar glaciers, air is trapped in bubbles that are ~8% of the total volume. The molar fraction gas content is thereafter fixed, with ratios between various gas species very similar to those of the atmosphere. If melt occurs subsequently, and the water begins to equilibrate with either the open atmosphere (e.g., if the melt occurs in the marginal apron) or bubbles having atmospheric composition, the differing solubilities of the gas species result in a gas composition of the liquid different from that of the atmosphere. After refreezing, gas bubbles in the refrozen melt will have ratios different from atmospheric. In particular, the Ar/N_2 ratio will be higher than atmospheric (Herron and Langway, 1979) due to the ~2 times greater water solubility of argon than nitrogen, and the CO_2 gas fraction will be higher than atmospheric due to the very high relative solubility of CO_2 (~70 times greater than that of N_2). Measurements of both Ar and CO_2 are complementary, because the former is inert whereas the latter has a very high solubility contrast. Furthermore, the total gas content will be substantially reduced relative to that for normal bubbly glacier ice.

Beneath Meserve Glacier, we expect the facies 3 gas composition to have a lower total gas content and increased Ar/N_2 and CO_2 relative to atmospheric, and it does (Fig. 2). The amber layer composition, however, is essentially identical to that of the overlying bubbly glacier ice, except near the facies 2–3 boundary, where physical mixing has combined the two (directly visible in the samples as finely stratified clear and amber ice). This is strong evidence that the amber layers were entrained without bulk melt and refreeze, the volume of interfacial films and grain boundary veins being too small to significantly alter the gas compositions (~0.001% in the most impure ices; Cuffey et al., 1999).

Stable Isotope Composition. The stable isotopic composition of the basal ices supports this view. Paired measurements of δD and $\delta^{18}O$ for ice samples from facies 1 and 2 are largely concordant (Fig. 3) with a line of slope 8, which is characteristic of precipitation (Craig, 1961) and of glacier ice unaltered by bulk melt and partial refreezing (Jouzel and Souchez, 1982; Souchez and Jouzel, 1984). Such alteration is likely under conditions of net melt, and deviations of basal ice isotopic composition from the meteoric relation have been observed at polar glaciers and interpreted as evidence of subglacial temperate conditions (Sugden et al., 1987; Souchez et al., 1988; Zdanowicz et al., 1996; Iverson and Souchez, 1996; Lorrain et al., 1999).

Summary. Because entrainment has occurred here without significant alteration of the bubbly glacier ice, we can rule out several mechanisms for genesis of these layers, including overriding of dirty marginal apron, basal freezing of ground water (Wilson, 1979), and shear folding of permafrost. More generally, it appears unlikely that entrainment occurred while the bed was at the bulk pressure-melting point. Partial loss of melt water would have occurred, driven by the steep surface slope of this glacier, altering the gas and isotopic compositions. In addition, water volumes would have been sufficient to affect isotopic composition during regelation, as shown by Iverson and Souchez (1996). Thus entrainment probably occurred here at subfreezing temperatures. Moreover, entrainment probably is occurring under modern conditions; many other glaciers in the Dry Valleys region have amber layers of this sort, indicating that the process is widespread and currently active.

Geomorphic Work of the Subfreezing Meserve Glacier

Accepting that the modern debris flux to the margins of Meserve Glacier is the result of entrainment at subfreezing temperatures, has subfreezing erosion been a geomorphologically significant process here? The Dry Valleys environment has been cold and dry for at least the past 10 m.y. (Denton et al., 1993). Assuming that the modern debris flux is representative of long-term rates, we can estimate how much erosion has occurred in this time by this process. Estimates of the total

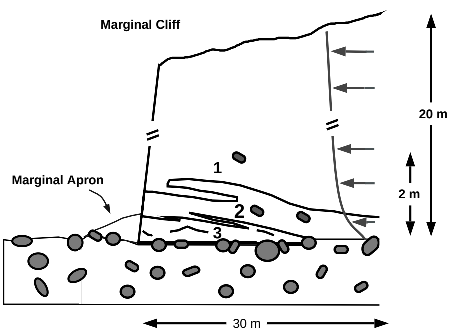

Figure 1. Vertical cross section through margin of Meserve Glacier, showing stratigraphy discussed by Holdsworth (1974) and reexposed by us. Ice flow is shown schematically by velocity vectors on right. Stratigraphy has three facies: 1, relatively clean, bubbly, ordinary glacier ice; 2, bubbly amber ice with dispersed fine rock particles; and 3, strained mixture of glacier ice blocks, snow, and refrozen meltwater, with stringers of particles, inferred to be marginal apron material overridden by glacier advance. At our tunnel site, glacier rests on bouldery till. Facies 3 tapers out at ~20 m from margin (Holdsworth and Bull, 1970). Physical mixing occurs by deformation at facies boundaries. Over time, facies 3 ice will be evacuated to margins by flow; current tapered-wedge geometry reflects relatively recent advance of ice margin, by at least 20 m.

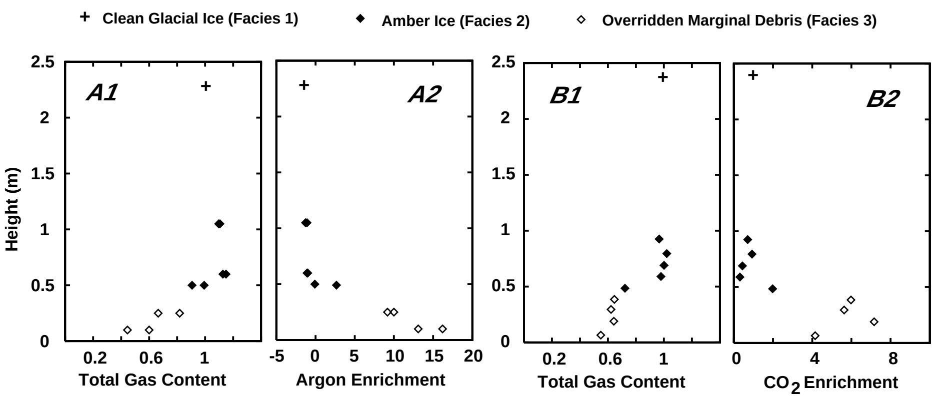

Figure 2. Gas composition measured on samples from two locations A and B in subglacial tunnel, at 9 and 11 m in from ice margin. A: Measurements by J. Severinghaus (there are two samples at each height). B: Measurements by R. Lorrain. Total gas content is normalized to that of top clean ice samples in A1 and B1. Argon enrichment for sample is defined as $\delta Ar/N_2$ (relative to atmospheric) normalized to $\delta Ar/N_2$ for top two samples in A2 (which look like one point). Value of $\delta Ar/N_2$ is negative in normal glacier ice because of size-dependent exclusion of Ar during bubble close-off. Measurements of $\delta Ar/N_2$ were corrected for gravitational separation in firn column. CO_2 enrichment is gas fraction of CO_2 normalized to that for top sample in B2. This has high CO_2 fraction (366 ppmv) relative to Holocene preindustrial level, which suggests some influence of summer melting in accumulation zone (Souchez et al., 1993).

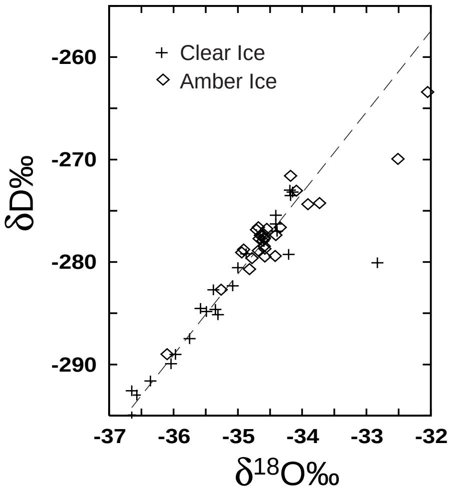

Figure 3. Coisotopic plot for amber ices and overlying clear glacial ice. Dashed line has slope of 8. Two anomalous samples from amber ice were taken from boundary with clear ice interlayers, and may have been entrained from underlying facies 3. Facies 3 ices have large ranges of δD (–295‰ to –220‰) and deviations from meteoric (to 20‰ in δD).

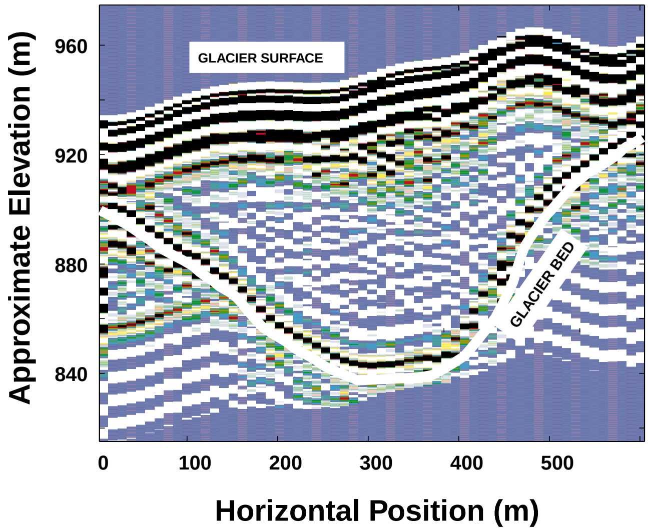

Figure 4. Radio-echo sounding image of glacier bed along east-west transect across upper part of ablation zone (lower ice tongue) corrected for surface elevation. Glacier bed (after data migration) is shown as thick white line. Records collected at 10 m spacing.

debris flux exuded by the Meserve Glacier tongue are 10 m^3 yr^{-1} to 30 m^3 yr^{-1} based on measurements of ice flux and debris concentration (Cuffey, 1999). The erosion rates corresponding to these two rock flux estimates are, as an average over the glacier bed, 9×10^{-7} to 3×10^{-6} m yr^{-1}, very low compared to those for temperate glaciers (Hallet et al., 1996). Over 10 m.y., the net erosion totals 10–30 m if averaged over the glacier bed.

Radio-echo sounding measurements show that the upper tongue of Meserve Glacier occupies a U-shaped trough, a classic landform of glacial erosion (Fig. 4), with a depth of ~60 m. The lower portion of the ablation zone is known to overlie a planar bed (Holdsworth, 1974), so the glacier has probably not eroded significantly here. In addition, the bed topography beneath the accumulation zone (data of John Clough, presented in Holdsworth, 1969, p. 124) shows that the concentrated glacial scouring only occurred in a central core region of the glacier bed. Thus the bed-averaged erosion of 10–30 m is an underestimate for the U-shaped trough, and the modern subfreezing entrainment rates are sufficient to account for much of this landform. Given the approximate nature of these estimates, the geomorphic significance of subfreezing vs. temperate processes is ambiguous; it is not necessary to interpret this landform as a relict of past temperate conditions, although the possibility is certainly not excluded.

Dry Valleys glaciers provide an extreme example in three respects. The elapsed time of cold-based conditions has been very long compared to that at most locations (Denton et al., 1993), and the solute content of the ice is high (~10^{-4} molar). However, temperatures here are very low; a –17 °C basal temperature is much lower than the basal temperatures of most thicker polar ice masses. Because erosion rates probably increase with basal temperature (see following), the possibility should be considered at other locations that erosion at subfreezing temperatures is a geomorphically significant process, and that landforms of glacial erosion beneath subfreezing ice are not relicts.

ROLE FOR WATER-DEPENDENT PROCESSES?

Entrainment of rock fragments beneath cold glaciers can probably occur by creep (Sugden and John, 1976), but the high viscosity of ice makes this mechanism a difficult one for fine-grained particles, like the silt and sand of the Meserve amber ices. Local melt and refreeze allowing slip at ice-rock interfaces can resolve this difficulty. This mechanism requires the presence of interfacial films, thin layers of water at the interfaces between ice and immersed solids such as rocks, that exist because the thermodynamic penalty for allowing the supercooled liquid to be present is more than offset by the reduction of interfacial free energy afforded by separating the two solid interfaces (e.g., Wettlaufer et al., 1996).

Traditionally, it has been almost universally accepted that sliding does not occur at ice-rock interfaces at temperatures lower than the bulk pressure-melting point (e.g., Sugden and John, 1976; Denton et al., 1993; Gow et al., 1997). The alternate view, that cold-based glaciers slide and ice negotiates small rock protrusions or particles by regelation, was argued on theoretical grounds by Shreve (1984) as an application of Gilpin's (1979) initial theory for interfacial films. There is now abundant laboratory evidence for the existence of interfacial films, including direct measurements of film thickness, to temperatures as low as –30 °C (Dash et al., 1995; Beaglehole and Wilson, 1994; Wettlaufer et al., 1996), and their existence in natural glacier ices should be assumed unless contrary evidence is found. Their glaciological manifestation as slip at the ice-rock interface has been observed at temperatures of –4 °C (Echelmeyer and Zhongxiang, 1987) and, during our investigations at Meserve Glacier, at the very low temperature of –17 °C (Cuffey et al., 1999). Abrasion, striation, and generation of fines can therefore occur beneath cold glaciers, contrary to common assumption (Summerfield, 1991; Gow et al., 1997), though at rates much smaller than those characteristic of warm beds. The rate should be small due to low slip rates and to small contact forces in the absence of downward ice flow associated with basal melt (Hallet, 1979). Striations have been observed on debris from cold glaciers (Mercer, 1971; Holdsworth, 1974). Likewise, films allow entrainment by regelation of cold ice past fine particles, driven by temperature or pressure gradients (Gilpin, 1979; Walder, 1986; Iverson and Semmens, 1995).

Slip rates (Shreve, 1984), and therefore the efficacy of associated entrainment mechanisms, increase with differential stress and increase strongly as interfacial film thickness increases. Primary physical controls on film thickness (Dash et al., 1995) are temperature, solute concentration, and atomic-scale surface properties. Thickness increases strongly as temperature increases toward the bulk melting point (Gilpin, 1979; Wettlaufer et al., 1996), and increases markedly with solute concentration if this concentration is high (Shreve, 1984; Beaglehole and Wilson, 1994; Wettlaufer, 1999). The great strength of the temperature dependence can be qualitatively appreciated from Shreve's (1984) theoretical calculation suggesting that slip rate decreases by two orders of magnitude between –1 and –10 °C. In cold glaciers, interfacial layers can have high solute concentrations even in relatively clean ice (Cuffey et al., 1999) because of solute rejection during regelation and the strong segregation of impurities to grain boundaries (Paren and Walker, 1971; Alley et al., 1986). Thus geomorphic activity beneath a subtemperate glacier is favored by a temperature close to the bulk melting point, and by a high impurity content. Entrainment sufficient to form compositionally distinct layers does not require a large total volume and therefore is not so constrained. Erosion rates will depend strongly on the nature of bed material, as is universally recognized.

CONCLUSIONS

Cold-based glaciers can actively entrain basal material. This provides a general mechanism for formation of dirty basal layers, and may in some cases be an important geomorphic process. Subfreezing entrainment is likely facilitated by interfacial water films, which also allow sliding and abrasion by cold glaciers. The common assumption that cold-based glaciers are "protective rather than erosional" (e.g., Denton et al., 1993, p. 168) is not true in the absolute sense, although is accurate relative to temperate glaciers.

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