Arctic, Antarctic, and Alpine Research, Vol. 42, No. 1, 2010, pp. 57–66

Sediment Melt-Migration Dynamics in Perennial Antarctic Lake Ice

*Steven M. Jepsen**
Edward E. Adams† and
John C. Priscu‡

*Corresponding author: Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91009, U.S.A.
Steven.M.Jepsen@jpl.nasa.gov
†Department of Civil Engineering, Montana State University, Bozeman, Montana 59717, U.S.A.
eda@coe.montana.edu
‡Department of Land Resources and Environmental Sciences, Montana State University, Bozeman, Montana 59717, U.S.A.
jpriscu@montana.edu

Abstract

We examined sediment melt-migration dynamics in the ice cover of Lake Fryxell, Taylor Valley, McMurdo Dry Valleys, Antarctica, using a combination of laboratory experiments, field observations, and modeling. The specific objectives were to determine the thermal conditions required for sediment melt and how sediment migration rates vary with meteorological forcings and ice microstructure. These characteristics are relevant to the influence of climate change on lake ice structure and ecosystem processes in polar regions. Sediment began melting through laboratory ice at -2 °C in simulated summer conditions, with warmer ice producing faster melt rates. An energy balance model, supported by our laboratory experiments, demonstrated that subsurface sediment can melt down to an equilibrium depth of ~2 m in two years. Field experiments and modeling revealed that surficial sediment melts at about half the rate of subsurface sediment because of heat losses to shallow, cold ice and the cold, dry atmosphere. Gravity flow of sediment along grain boundaries was pronounced in laboratory ice warmer than -1 °C. This mechanism produced a flux of 0.1 g m^{-2} hr^{-1}, a significant value relative to published benthic sedimentation rates for these lakes indicating an important sediment sorting mechanism.

DOI: 10.1657/1938-4246-42.1.57

Introduction

The McMurdo Dry Valleys (dry valleys) is the largest (~4000 km^2) ice-free area of Antarctica (Bull, 1966) and the subject of extensive research on ecosystem processes and regional climate change (Priscu, 1998; Doran et al., 2002a). The physical conditions of biological habitats in the dry valleys are among Earth's most extreme (e.g., Ugolini, 1970; Friedmann, 1982; Priscu et al., 2005). A striking feature of this region is the presence of perennially ice-covered lakes that provide an oasis for life in this polar desert (Priscu et al., 1998, 2008).

Microbial consortia in ice covers of the dry valley lakes are closely associated with sediment (Priscu et al., 1998, 2005; Paerl and Priscu, 1998; Fritsen and Priscu, 1998), which is a primary location for liquid water 2–3 months per year (Fritsen et al., 1998; Adams et al., 1998). In addition to this ice-bound ecosystem, the sub-ice environments include benthic mats thought to be analogues of Precambrian stromatolites (Parker et al., 1981; Wharton et al., 1982) and plankton adapted to extremes in temperature and sunlight (Lizotte and Priscu, 1992; Morgan-Kiss et al., 2006). Lake properties affected by sediment transport include water column turbidity and primary productivity (Foreman et al., 2004), water column nutrient enrichment (Dore and Priscu, 2001), availability of organic compounds for benthic heterotrophs (Simmons et al., 1986), and the flux of microorganisms through lake ice to the water column (Priscu et al., 1998). Meltwater production around sediment in ice is also relevant to potential habitable conditions on early Mars (McKay, 1986; Doran et al., 1998; McKay et al., 2005), when ice-covered lakes may have existed in the Valles Marineris (Lucchitta and Ferguson, 1983; Squyres, 1989). Sediments also sorb hydrocarbons (e.g., Fetter, 1999), providing a potential dispersal agent of hydrocarbon contaminants in lake ice (Jaraula et al, 2008).

Lake ice sediment and glacier cryoconites (Sharp, 1949, and references therein) share similar thermodynamic processes. The depth and shape of cryoconites are an indicator of energy fluxes near the glacier surface (Gribbon, 1979; McIntyre, 1984; Wharton et al., 1985). Similarly, the location of sediment layers in perennial lake ice are an indicator of the energy fluxes (McKay et al., 1985; Simmons et al., 1986; Wharton et al., 1993) which are relevant to the mass and energy balance of the Taylor Valley lakes (Simmons et al., 1987; Wharton et al., 1992; Barrett et al., 2008).

We present results from field and laboratory experiments and a numerical model on the melt-migration of sediment in the ice cover of Lake Fryxell, Taylor Valley, to examine (i) the thermal conditions required for sediment melt, (ii) how sediment migration rates vary with meteorological forcings, and (iii) how ice microstructure influences sediment migration. The numerical model follows from previous work (Gribbon, 1979; McIntyre, 1984; Wharton et al., 1985) with assumptions applicable to subsurface conditions.

Study Area

Lake Fryxell is located in lower Taylor Valley of the dry valleys, Antarctica, 6.2 km from McMurdo Sound. The lake is situated 16 m above sea level, with a length, width, and maximum depth of 5.2 km, 1.2 km, and 20 m, respectively (Wagner et al., 2006). The mean annual air temperature at Lake Fryxell is approximately -20 °C, which is 2–3 °C colder than at neighboring Lakes Hoare and Bonney further west (Doran et al., 2002b). The summer air temperatures at Lake Fryxell are subfreezing, with mean values ranging from -5 to -8 °C (Doran et al., 2002b; Fountain et al., 1999). The dry valleys receive less than 10 cm of water-equivalent precipitation annually in the form of snow (Bull, 1966; Thompson et al., 1971; Bromley, 1985). Hydrologic recharge

1523-0430/10 $7.00

of Lake Fryxell is limited to seasonal inflow from glacial meltwater streams, and there are no outflowing streams (Conovitz et al., 1998; Chinn, 1993; McKnight et al., 1998). The water temperatures in Lake Fryxell increase from 0 °C at the ice-water interface to ~3 °C at the bottom (20 m), reaching a maximum of ~3.5 °C at mid-depth (Spigel and Priscu, 1998). The ice cover blocks about 97% of incident Photosynthetically Active Radiation (PAR, 400–700 nm) from entering the lake during periods when solar radiation is available (Howard-Williams et al., 1998; Parker et al., 1982; Wharton et al., 1982).

Ice cover persistence on the dry valley lakes results from a balance between ablation, freezing, and glacial meltwater recharge (Wilson, 1981; McKay et al., 1985). Summer air temperatures must be high enough that glaciers melt and recharge the lakes, otherwise the lakes would disappear as the result of sublimation and evaporation. At the same time, air temperatures must allow for a balance between the accretion of new (bottom) ice and the ablation of old (surface) ice. The ablation rate of the Lake Fryxell ice cover varies considerably because of summertime melt and the dry katabatic winds which are most frequent during non-summer months (Henderson et al., 1966; Doran et al., 2002b). The ablation rates at Lake Fryxell ranged from 30 to 40 cm yr^{-1} from 1961 to 1963 (Henderson et al., 1966), and from 34 to 51 cm yr^{-1} from 2001 to 2004 (MCM LTER). These rates are similar to the rate of 35 cm yr^{-1} at Lake Hoare in 1986 (Clow et al., 1988). Much higher ablation rates, up to ~1.5 m yr^{-1}, were apparent at Lake Hoare from 1983 to 1985 (Simmons et al., 1986). Based on MCM LTER records for 2001–2004, the surface ablation rates at Lake Fryxell during the months of December and January were an order of magnitude greater than during other months. Intraseasonal variation in surface ablation was also noted at Lake Hoare by Clow et al. (1988) and is likely associated with summertime melting at the ice surface. The ice cover thickness of Lake Fryxell ranged from 3.8 to 4.5 m between 1961 and 1963 (Henderson et al., 1966), and from 3.5 to 6.0 m between 1995 and 2006 (Priscu, unpublished data; MCM LTER).

Much of the lake floor sediment in the dry valleys originated as eolian deposits that migrated through the ice covers by melting and settling along fractures and conduits (Adams et al., 1998; Simmons et al., 1986; Nedell et al., 1987; Squyres et al., 1991). Lancaster (2002) reported an eolian sediment flux of 1 g m^{-2} yr^{-1} on the ice of Lake Fryxell, of which 89% (by mass) was mud and 11% was sand. The sand fraction may have been underestimated owing to the elevated location of the collection pans (Lancaster, 2002).

Methods

LABORATORY EXPERIMENTS

Five experiments (tests 1–5) measuring sediment melt-migration in ice were conducted at the SubZero Science and Engineering Laboratory at Montana State University, Bozeman (http://www.coe.montana.edu/ce/subzero/). Ice blocks containing sediment were exposed to Lake Fryxell meteorological conditions in an environmental chamber equipped with a metal halide lamp. The ice blocks measured 25 × 31 cm horizontally and 25, 16, 17, 33, and 49 cm vertically in tests 1–5, respectively. In tests 2–5, the sediment was placed in a cylinder-shaped melt cavity, measuring 5 cm horizontally and 2.5 cm vertically, in ice that was insulated to reduce horizontal temperature gradients. In test 1, the sediment was placed in a 1-mm-thick layer over the ice surface. This ice was not insulated to prevent smearing of mud. Before each experiment, the ice and sediment were allowed separately to equilibrate with the chamber atmosphere. Shortwave radiation was then activated and sediment was added to the ice (zero hours defined) and allowed to melt downward into an underlying collection pan. A shortwave irradiance of 589 ± 11 W m^{-2} (285–2800 nm), measured with an Eppley PSP pyranometer, and air temperature of −5.1 °C were applied to represent mid-November, peak-sunlight conditions at Lake Fryxell. Simulations of meteorological conditions occurring earlier in the year revealed no melting of ice with sediment. The humidity and independently controlled ceiling temperature of the environmental chamber were maintained at 30 ± 3% and −1.5 ± 0.7 °C. Temperatures were recorded in the sediment and three vertical locations in the ice using Type-T thermocouples (0.2 °C accuracy) and a Campbell datalogger. The ice temperature at the vertical level of the sediment was estimated by linearly interpolating values between thermocouples. The sediment depths were measured relative to the initial ice surface in order to compute absolute velocities, requiring knowledge of upper surface ablation. We measured the total vertical ablation while noting significant melt at both the upper and lower surfaces. We therefore assumed that the upper surface ablation was half of the total and carried the uncertainty through our velocity calculations assuming random propagation of error (e.g., Preston and Dietz, 1991).

The sediment of tests 2–5 (not test 1) was 50% saturated with JP8 jet fuel (0.1 mL g^{-1} sediment) as part of a separate contaminant transport study (hydrocarbon data not included in this paper). Migration rates of uncontaminated and contaminated sediment in ice were compared in a separate experiment, revealing no discernible difference in ice temperature at incipient melt, and an insignificant (3%) difference in sediment depths by the end of the experiment (30 hrs). Hence, JP8 at 50% saturation was found to have an insignificant effect on sediment melt-migration. The sediment in all tests was collected during transects across Lake Fryxell in 1996. The mass fractions were 4% mud (<0.06 mm), 41% fine sand (0.06–0.3 mm), and 55% coarse sand (0.3–2.0 mm); gravel (>2 mm) was excluded. The laboratory ice was clear, unfractured, consisted of crystals measuring 0.5–7 cm horizontally and 9–28 cm vertically, and was grown directionally upward in a vessel with circulating water. The crystal *c*-axes were horizontally aligned, unlike the vertical alignments typical at Lake Bonney and presumably the other dry valley lakes.

FIELD EXPERIMENTS

The fieldwork portion of this study was conducted between November 2007 and March 2008 at Lakes Fryxell and Bonney in Taylor Valley. On 10 November 2007, sediment was collected from the north shore of Lake Fryxell, sieved to remove grain sizes exceeding 4 mm, and placed on the perennial ice in a cm-thick layer covering a 0.9 × 1.2 m area (Fig. 1). The depth of this "sediment patch" was measured relative to the surrounding, sediment-free ice surface once every 1–3 weeks until 17 January 2008. Downwelling PAR was measured on the ice surface, next to the sediment, with a Licor 190-SA quantum detector and LI-1000 datalogger. These values were converted to shortwave irradiance using a conversion factor of 0.5 W m^{-2} per µmol photon m^{-2} s^{-1} (Clow et al, 1988). Albedos of beach sand and ice table (Fig. 1) were measured at nearby Lake Bonney (Lake Fryxell values unavailable) on 26 March 2008 using a TriOS RAMSES-ACC hyperspectral radiometer, cosine-corrected to 6–10% accuracy. The beach sand was frozen and visibly resembled the lake ice sediment. A single scan of upwelling and downwelling PAR (320–950 nm) was obtained at each facies, integrated over wavelength,

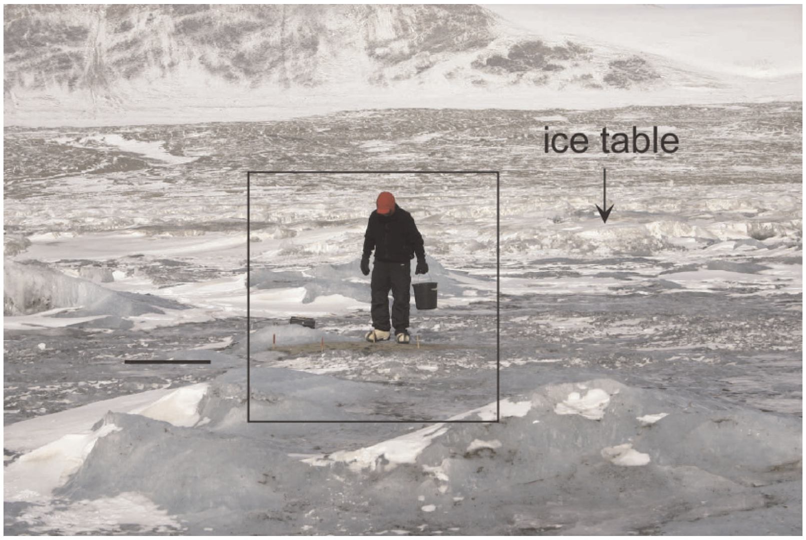

FIGURE 1. Looking south over Lake Fryxell at the beginning of the sediment melt experiment, 10 November 2007, scale bar 1 m. The introduced sediment is below the person in the box. The topography was relatively flat near the experiment, with "ice tables," 0.5–1 m in height, visible in background.

and divided to obtain all-wave albedos. Ice cover thicknesses and depths to piezometric level were measured at Lake Fryxell from drill holes placed within 20 m of the sediment patch. In early November 2007, the lake ice around the sediment patch was relatively flat and light blue in color (Fig. 1). Sparsely scattered sediments were visible in the upper ~2 cm of ice, some resembling the "dust wells" shown in Sharp (1949). White, heavily weathered ice tables, 0.5–1 m in height, were abundant at distances greater than ~10 m from the sediment patch (Fig. 1).

NUMERICAL MODEL

Sediment melt through ice was modeled using steady state energy conservation on a control volume of sediment. Shortwave energy incident to the sediment is absorbed because of its low albedo, warming the sediment and potentially causing melt. The net energy flux to the sediment provides the latent heat for melt (McIntyre, 1984; Wharton et al., 1985):

$$\rho_i L v = Q_I - \Sigma Q_{out} \tag{1}$$

The terms ρ_i, L, and v are the ice density (917 kg m^{-3}), heat of fusion for water (3.34×10^5 J kg^{-1}), and absolute sediment velocity, respectively (Table 1). The latter is given by $v_s + dz/dt$, where v_s is the surface ablation rate, z is the sediment depth, and t is time (Wharton et al., 1985). The left side of Equation 1 represents the latent heat flux of melt, Q_I is the shortwave energy absorbed by the sediment, and ΣQ_{out} is the total outward flux from thermal conduction, sensible and latent heat loss to air, and longwave radiation. Our objective was to compute the melt rate of sediment after it has melted through the surface and become ensconced in ice. We therefore assumed that the outward energy fluxes were limited to thermal conduction from the sediment, Q_C. The energy balance relation simplified to:

$$\rho_i L v = Q_I - Q_C \tag{2}$$

Thermal conductive flux from the sediment is proportional to the temperature gradient in ice at the boundary of the sediment. To determine this, we assumed that the temperature gradient is proportional to the difference between the sediment surface temperature and the ice temperature, T_{ice} (z,t), an infinite horizontal distance away (Simmons et al., 1986), the latter of which we associated with the ice temperatures measured in our laboratory experiments. For the case of sediment melting through ice, thermal conductive flux scales with ice temperature:

$$Q_C = -K\lambda T_{ice}(z,t), \quad \text{(boundary of sediment at 0 °C)} \tag{3}$$

where K is a thermal conduction coefficient (length^{-1}) and λ is the thermal conductivity of ice (2.14 W m^{-1} °C^{-1}). We determined the value of K by comparing the sediment melt rates and ice temperatures measured in the laboratory (see results section).

The absorption of shortwave energy per unit area of sediment was modeled using the Beer-Lambert law:

$$Q_I = T(1-R)(1+S)I_0(t)\exp(-kz), \tag{4}$$

where T is the fraction of downwelling shortwave radiation transmitted into the ice through a semi-opaque surface layer, such as snow, R is the all-wave albedo of sediment, S is the ratio of upwelling to downwelling shortwave flux at the sediment, $I_0(t)$ is the downwelling shortwave flux above the ice surface, and k is the all-wave absorption coefficient for ice. The value of R was determined by matching Equations 2 and 4 to the laboratory-

TABLE 1

Notation.

h	Vertical distance from perched ice table to surface sediment
I_0	Downwelling shortwave irradiance at ice surface
K	Empirically derived thermal conduction coefficient (length^{-1})
L	Heat of fusion for water, 3.34×10^5 J kg^{-1}
MCM LTER	McMurdo Dry Valleys Long Term Ecological Research
PAR	Photosynthetically Active Radiation (400–700 nm)
Q_C	Flux of thermal conduction from sediment
Q_I	Flux of shortwave radiation to sediment
R	Sediment all-wave albedo
S	Ratio of upwelling to downwelling shortwave irradiance
T	Fraction of shortwave radiation transmitted through a semi-opaque surface layer
T_{ice}	Ice temperature
t	Time
v	Absolute sediment velocity
v_s	Ablation rate of ice surface
z	Depth of sediment in ice
α	Fraction of ice surface covered by sediment
λ	Thermal conductivity of ice, 2.14 W m^{-1} °C^{-1}
ρ_i	Density of ice, 917 kg m^{-3}
ρ_w	Density of water, 10^3 kg m^{-3}

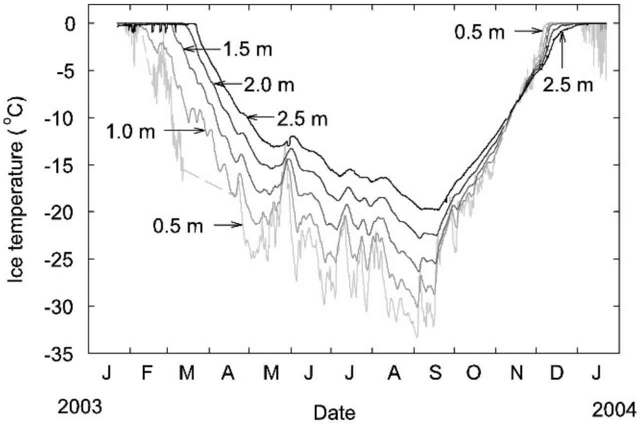

FIGURE 2. Lake Fryxell ice temperatures from January 2003–January 2004. Dashed lines indicate missing data replaced using interpolation. The data were adjusted for a measured surface ablation rate of 12 cm yr^{-1}.

measured sediment velocities in ice at 0 °C (see results section). We tested T values of 1.0, 0.52, 0.25, and 0.15 to simulate surface layers of 0.0, 0.5, 1.0, and 2.0 mm liquid-equivalent fresh snow, respectively. These values are based on findings by Wiscombe and Warren (1980) that a dark surface covered with 1 and 2 mm of liquid-equivalent fresh snow has an albedo of 0.75 and 0.85, respectively. The T value for a 0.5 mm thickness was found using interpolation. We used constant values of k and S, although these vary with depth and wavelength (McKay et al., 1994). We tested k values ranging from 1.0 to 1.21 m^{-1} based on measurements of the Lake Fryxell ice cover by Howard-Williams et al. (1998). Typical measured ratios of upwelling to downwelling shortwave, S, over Lake Fryxell during the months of December and January range from 0.2 to 0.3 (MCM LTER). McKay et al. (1994) reported S values ranging from 0.3 to 0.5 in the upper 2.5 m of Lake Hoare ice. Recognizing that these values will vary with location on the ice, we set S to a reference value of 0.25 and tested other values indirectly by varying the parameter T.

Equations 2–4 form a first-order nonlinear differential equation for $z(t)$. In ice warmer than the "melt-threshold" temperature of $-Q_I\,(K\lambda)^{-1}$, sediment melt rate increases linearly with ice temperature, with rates vanishing below this threshold. The depth of a sediment layer in the Lake Fryxell ice cover between 1 November 2003 and 1 November 2007 was predicted by integrating Equations 2–4 using forward differencing and a time step of 0.05 days. The model was forced with shortwave radiation and lake ice temperatures (MCM LTER) measured at Lake Fryxell in 2003 (Fig. 2). The same forcings were used for each year modeled. The sediment depth was initialized to 0.5 m, the shallowest depth of the temperature profile owing to sensor meltout from ablation (Fig. 2).

The energy balance relation of our model differs in two ways from the relation for cryoconites in McIntyre (1984). First, our model is for subsurface sediment and therefore excludes longwave radiation and losses to the atmosphere in the form of sensible and latent heat. Second, we considered the conductive energy losses from sediment in different thermal conditions (Equation 3), allowing us to predict the seasonal timing of meltwater generation around sediment. In addition, our model was tested in a laboratory setting using observed sediment melt rates, providing a data set to accompany those based on field measurements in previous studies (Gribbon, 1979; Fountain et al., 2004).

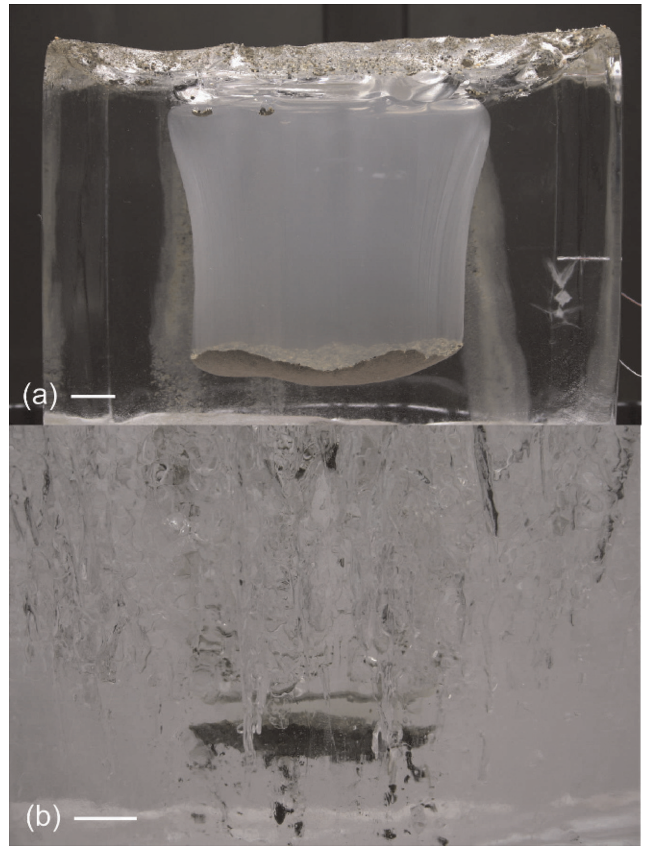

FIGURE 3. Sediment layers in laboratory ice. (a) Test 1, 48 hrs elapsed: sediment thickness had increased from 1 mm to ~1 cm (shown), scale bar 3 cm. (b) Test 5, 102 hrs elapsed: sediment thickness had decreased from 2.6 cm to 1.3 cm (shown), scale bar 2 cm.

Results

LABORATORY RESULTS

The sediment evolved into layers ~1 cm thick during all experiments. During test 1, the sediment layer thickened from 0.1 to 1 cm (Fig. 3a). During test 5, the sediment layer thinned from 2.6 to 1.3 cm (Fig. 3b). These dimensions are characteristic of sediment aggregates occurring in lake ice of the dry valleys (Adams et al., 1998; see also Priscu et al., 1998). The melt tube in test 1 backfilled with meltwater (Fig. 3a), while the tubes in the other tests drained via water infiltration through ice. We attributed this to the ~1 °C lower temperature of ice in test 1 which was uninsulated.

The main sediment disks exited the ice at elapsed times of 60, 20, 20, 35, and 110 hrs in tests 1–5, respectively. At earlier times, mud and fine sand settled downward in meltwater along grain boundaries of ice warmer than −1 °C. At an elapsed time of 8.5 hrs in test 5, sediment was observed traveling downward along a 2-mm-diameter conduit (Fig. 4a). Based on the location of internal melt figures, we concluded that this conduit followed a grain boundary junction (Fig. 4a). At this time, the temperature near the top of the ice was −1 °C and sediments were observed as deep as 17 cm, indicating an average velocity of ~2 cm hr^{-1}. In an experiment identical to test 2, but using an ice block 25 cm in height, a coating of mud and fine sand was observed underneath the ice when the main sediment disk was 15 cm deep (Fig. 4b). Sediments in this coating were concentrated along heavily melted grain boundary grooves, imparting a polygonal pattern. From photos taken during a 12 hr period in test 4, about 4 mg of sand

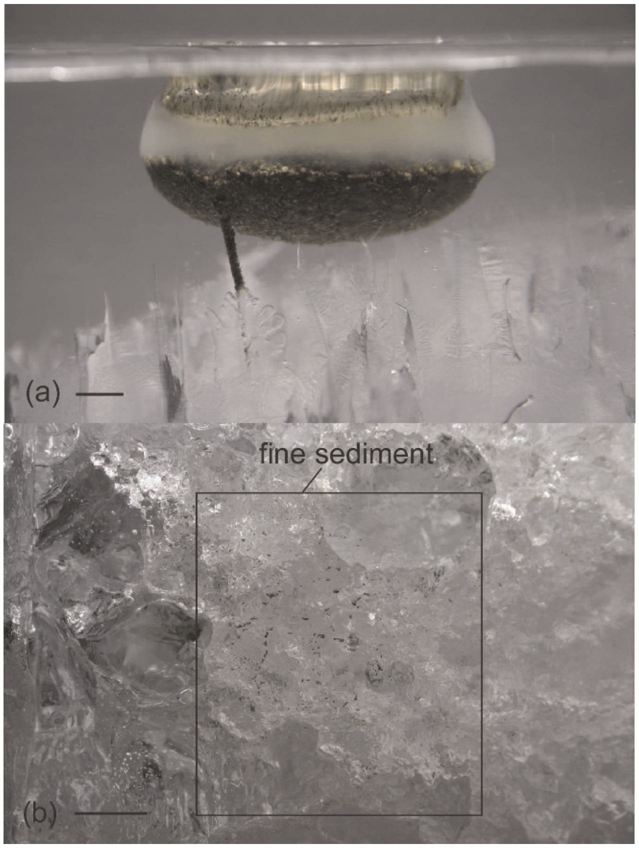

FIGURE 4. Structural features in ice accommodating intergranular sediment flux. (a) Test 5, 8.5 hrs elapsed: sediment traveling downward in a 2 mm diameter conduit along a grain boundary junction, scale bar 1 cm. The ice temperature was approximately −1 °C. (b) Bottom of an ice block showing fine sediment concentrated in heavily melted grain boundary grooves, scale bar 2 cm (from preliminary experiment). At the time of this photo, the main sediment layer had melted through 60% of the ice.

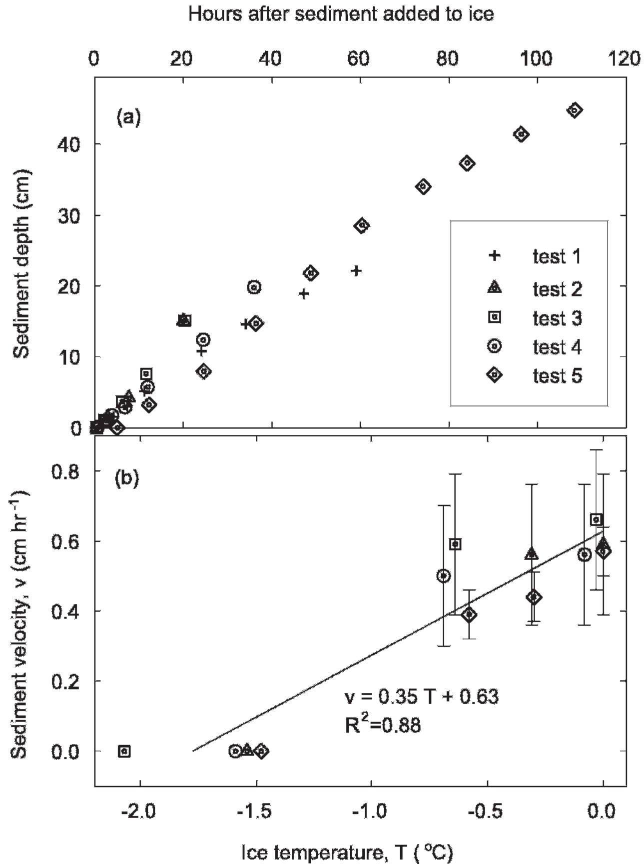

FIGURE 5. Laboratory measured (a) sediment depth (±0.7 cm) versus time, and (b) absolute sediment melt velocity versus ice temperature. Sediment depth is measured from the original ice surface. Ice temperatures shown are at the depth of the sediment.

(~60–150 μm grain size) passed along grain boundaries into the collection pan, yielding a flux of ~0.1 g m^{-2} hr^{-1}. This flux was also found by weighing of sediments in meltwater collected during test 5.

Sediment depths relative to the initial ice surface are shown in Figure 5a. The uncertainties ranged from 0.2 to 1.7 cm (0.7 cm average). About half of this was from uncertainty in surface ablation, the other half from parallax. The sediment temperatures increased to a maximum of 2–3 °C during the first 5–8 hrs of the experiments. Soon thereafter, the sediment thermocouples fell out owing to sediment disaggregation. The ice in tests 2–5 gradually warmed from initial temperatures between −2.5 and −4.5 °C to the melting point by elapsed times ranging from 5 to 8 hrs. Liquid water was first noticed around sediment in ice at −2 °C and indicated the zero velocities shown in Figure 5b. As the ice approached the melting point, the absolute sediment velocities noticeably increased to a maximum of ~0.6 cm hr^{-1} (Fig. 5b). This increase presumably resulted from decreasing thermal conduction from sediment in warmer ice.

The velocities in Figure 5b are from the first half of the experiments for two reasons. (1) In the second half of the experiments, internal melting in ice was accompanied noticeably by light scattering. This was also apparent from sediment velocities exceeding the upper limit of $Q_I\,(\rho_i L)^{-1} = 0.69$ cm hr^{-1} for the case of zero irradiance of scattered light (Equation 2). We did not use sediment velocities during these times because we measured only the direct beam irradiance. (2) The quality of thermal contact between the thermocouples and ice progressively degraded due to increasing water saturation.

FIELD RESULTS

The all-wave albedos of beach sand and ice table at Lake Bonney were 0.14 and 0.48, respectively. This sand albedo is close to the average value of 0.13 (400–700 nm) measured at Lake Hoare by McKay et al. (1994). The introduced sediment patch on Lake Fryxell melted downward, while remaining on the surface (Fig. 6a), at an average rate of 1.2 cm day^{-1} relative to the surrounding ice (Fig. 7). Because the surrounding ice also ablated at an estimated rate of 0.1 cm day^{-1} (Henderson et al., 1966; Clow et al., 1988), the absolute velocity of the sediment patch was about 1.3 cm day^{-1}. The average downwelling shortwave measured near the sediment was 310 W m^{-2}, a value which was 6.7 times greater than the latent heat of meltwater generation below the sediment patch (Equation 1). Ice layers, 1–20 mm in thickness and consisting presumably of refrozen meltwater, were observed over the sediment patch. These layers collapsed between visits or during data collection. Surficial sediment was often observed forming clumps set into the ice along crystal interstices during January 2008 (Fig. 6b).

The Lake Fryxell ice cover was 4.6 m in thickness during November and December 2007. Drill holes in the upper meter of lake ice began backfilling with water sometime between 7 and 27 November 2007, signaling an increase in ice permeability. The

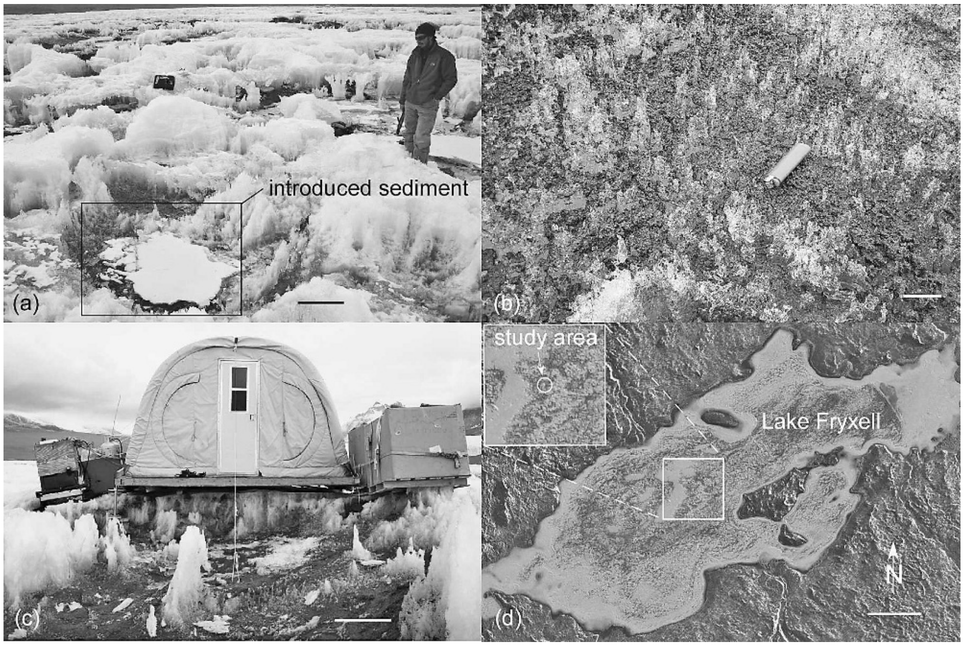

FIGURE 6. (a) South-south-east view over Lake Fryxell on 8 January 2008, scale bar 0.5 m. The introduced sediment (boxed) was similar in appearance to other naturally occurring sediment patches. (b) Sediment in clumps set into the ice along crystal interstices, 8 January 2008, scale bar 5 cm. (c) Looking west at a sampling hut on 17 January 2008, scale bar 1 m. The hut floor was level with the surrounding surface in early November 2007. (d) Quickbird 60-cm-resolution photo of Lake Fryxell, 16 March 2007, showing sediment patches and the location of our sediment melt experiment at 77.6105°S, 163.1468°E, scale bar 0.5 km (map from Antarctic Geospatial Information Center).

measured depths to piezometric level in the ice cover increased from 0.4 m in late November 2007 to 1.1 m in mid-January 2008, changing abruptly in late December (Fig. 7).

Between November 2007 and January 2008, the topography of the site changed from relatively flat (Fig. 1) to ~1 m in relief, prompting relocation of the sampling hut (Fig. 6c). The topography in January consisted of heavily weathered ice tables (albedo ~0.48; Figs. 6a, 6c) perched ~1 m above areas covered in sediment. This caused some difficulty in distinguishing the introduced sediment patch from surrounding, naturally formed patches (Fig. 6a). A satellite view of these ice tables and sediment patches at the end of summer 2007 is shown in Figure 6d.

MODEL RESULTS

We determined the remaining model parameters, which are the sediment albedo (R) and thermal conduction coefficient (K), using data measured in the laboratory experiments. The sediment albedo was determined by comparing Equations 2–4 with the observed sediment velocity of ~0.6 cm hr^{-1} in ice at the melting point. Because light scattering to the sediment appeared minor, we assumed that $S = 0$ and computed a sediment albedo of 0.1 for use in the model. We also modeled other albedos indirectly by varying the parameter T. We computed a thermal conduction coefficient of 140 m^{-1} by matching Equations 2–4 to the linear regression in Figure 5b. To test other K values, we sought the first-order dependence of K. Thermal conductive flux from individual sediment particles has been suggested to scale inversely with grain size (Simmons et al., 1986). Applying this argument to aggregated sediment, we assumed that K scales inversely with sediment layer thickness and tested K values of 140, 280, and 560 m^{-1} for layer thicknesses of 2, 1, and 0.5 cm, respectively.

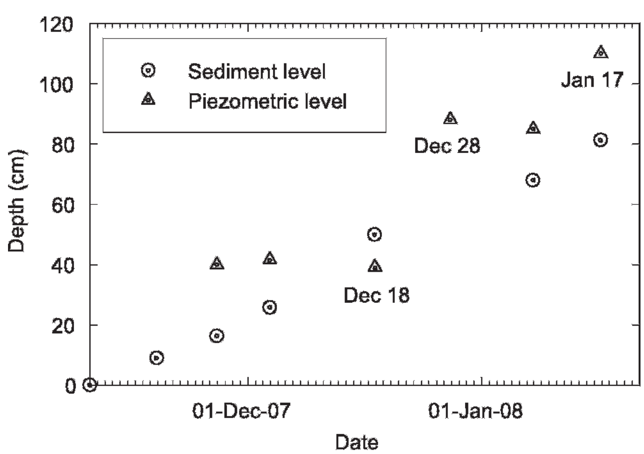

FIGURE 7. Depth of sediment and piezometric level relative to the sediment-free ice surface, Lake Fryxell. The slope of the sediment curve is 1.2 cm day^{-1}. The piezometric depth increased abruptly from 0.4 to 0.9 m in late December 2007.

The modeled depths of a sediment layer in the ice cover of Lake Fryxell are shown in Figure 8. Each figure panel shows model output with variation of a single parameter. The solid curves show results for the reference parameter set $K = 140$ m^{-1}, $k = 1.08$ m^{-1}, $v_s = 0.35$ m yr^{-1}, and $T = 1$ (no surface snow). For these parameters, the sediment layer reaches an equilibrium depth, where average melt and ablation rates become equal, of ~2.2 m in two melt seasons. Most of this depth (90%) is reached after one summer. The modeled depths are most sensitive to surface snow thickness (Fig. 8a) and ice ablation rate (Fig. 8b). A thin layer of surface snow greatly reduces the equilibrium depth (Fig. 8a). Greater ice ablation rates have the effect of decreasing the average sediment depth while increasing its temporal variation (Fig. 8b). The modeled depths are relatively insensitive to the range in k and K that are likely to occur in the lake ice (Figs. 8c, 8d). The model results are also relatively insensitive to possible error in sediment albedo. Increasing R from 0.10 to 0.14 is equivalent to decreasing T from 1.0 to 0.96 (Equation 4), resulting in only an ~2% reduction in equilibrium depth (Fig. 8a).

Discussion

Our model and laboratory experiments demonstrate that a balance between shortwave absorption, thermal conduction, and latent heat can account for the downward melt of sediment to an equilibrium depth of ~2 m in dry valley lake ice. Transport to

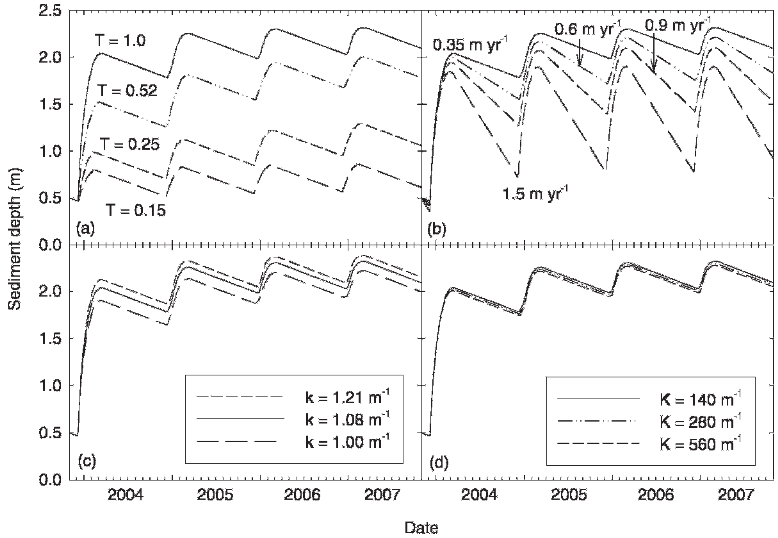

FIGURE 8. Modeled depth versus time of a sediment layer in the ice cover of Lake Fryxell. Each plot shows model output with variation of a single parameter. *T* values of 1.0, 0.52, 0.25, and 0.15 simulate surface snow layers of 0, 0.5, 1.0, and 2.0 mm liquid-equivalent, respectively. v_s = ice surface ablation rate; *K* = thermal conduction coefficient, values of 560, 280, and 140 m^{-1} represent sediment layer thicknesses of 0.5, 1, and 2 cm, respectively; *k* = all-wave absorption coefficient of ice. Unless noted otherwise, $T = 1.0$, $v_s = 0.35$ m yr^{-1}, $k = 1.08$ m^{-1}, and $K = 140$ m^{-1}.

deeper ice requires the presence of fractures or conduits (Simmons et al., 1986; Nedell et al., 1987; Squyres et al., 1991). An equilibrium depth of 2 m compares well with published depths of accumulated sediment in the ice covers of Lakes Bonney (Priscu et al., 1998; Fritsen and Priscu, 1998) and Hoare (Nedell et al., 1987; Squyres et al., 1991). However, this depth exceeds values observed at Lake Fryxell, where sediment accumulates in the upper 0.5 m of ice (Fritsen and Priscu, 1998; Priscu et al., 2005). Priscu et al. (2005) attributed relatively shallow ice sediment at Lake Fryxell to the higher snowfall rates in closer proximity to McMurdo Sound, a hypothesis which is consistent with results from the model (Fig. 8a). Another reason for the shallow ice sediment at Lake Fryxell may be the 2–3 °C lower air temperatures than at Lakes Hoare and Bonney (Doran et al., 2002b), causing colder ice and thus lower sediment melt rates.

The sediment in our lab and field experiments formed layers and clumps with characteristic dimensions of 1–2 cm. These dimensions are similar to those in the ice covers of Lakes Bonney (Adams et al., 1998; Priscu et al., 1998) and Hoare (Nedell et al., 1987; Squyres et al., 1991). Sediment coalescence into clumps and layers may enhance melt-migration by increasing shortwave absorption and thermally insulating grains from ice. This may account for the discrepancy between studies showing that sediment melts effectively through lake ice (this study; Adams et al., 1998; Fritsen et al., 1998) and studies suggesting that sediment melt-migration is inhibited by thermal contact between ice and individual grains (Simmons et al., 1986; Nedell et al, 1987; Wharton et al., 1993).

The Lake Fryxell sediment patch remained on the ice surface during melt, rather than becoming enclosed in ice. Two potential mechanisms by which surficial sediment becomes enclosed in ice are: (1) downward migration along crystal interstices (Fig. 6b; Sharp, 1949), and (2) lake water infilling and subsequent freeze-over in depressed areas. Indeed, we observed a piezometric level above the sediment patch on 18 December 2007; however, the water line receded before freezing around the sediment (Fig. 7).

The total vertical ablation of the Lake Fryxell sediment patch, 80 cm (Fig. 7), was twofold less than the modeled result (Fig. 8). We identified two possible causes for this discrepancy while noting that the model was not intended for surface applications. First, energy loss to the atmosphere was not modeled but is significant for surficial sediment (Drewry, 1972; McIntyre, 1984; Hendy, 2000). Second, ice temperatures increase with depth during 10 months of the year (Fig. 2; Fritsen and Priscu, 1998), resulting in less sensible heat loss from sediment with depth. The model results were insensitive to the thermal conduction coefficient, *K*, because the lake ice temperatures rapidly transition between the melt-threshold and 0 °C (several weeks, Fig. 2), while sediment melt takes place over a longer period of 2–3 months (Fig. 8; Fritsen et al., 1998; Adams et al., 1998).

We identified four possible causes for the difference between a computed sediment albedo of 0.1 and a value of 0.14 measured at Lake Bonney: (1) differences in water saturation or biological composition led to a difference in reflectance; (2) unmeasured light scattering in the laboratory resulted in a higher actual sediment irradiance, causing us to underestimate the sediment albedo; (3) the 0.2 °C uncertainty in measured ice temperature translated to an albedo uncertainty of 0.1 (Equations 2–4); and (4) the assumption of negligible losses to the laboratory atmosphere led to an underestimate in modeled energy flux from the sediment. Including these fluxes would have decreased the computed sediment albedo for any given melt rate (Equation 2 with T_{ice} = 0 °C). We believe that this was not the main source of error since our computed sediment albedo already seemed too low.

Ice can act as a filter by preferentially transporting fine sediments. Two different mechanisms can explain this. One, finer sediment may have an enhanced ability to melt ice (Hendy, 2000). Two, finer sediment can preferentially pass between intergranular pores. We observed the latter mechanism in our laboratory experiments, where fine sediment traveled through intergranular ice pores in meltwater generated from the "solid-state greenhouse" effect (Schlatter, 1972; Brandt and Warren, 1993). The intergranular sand flux of 0.1 g m^{-2} hr^{-1} that we measured translates to a 100 g m^{-2} yr^{-1} flux assuming a two month melt season. This flux will vary depending on sediment quantity, grain size, and ice microstructure. The flux we measured is within a factor of 3 of average sediment fall-out rates from ice at Lake Hoare (29–41 g m^{-2} yr^{-1} at 3/4 sites) (Simmons et al., 1986; Nedell et al., 1987; Squyres et al., 1991), and far exceeds an eolian sediment flux of 1 g m^{-2} yr^{-1} on Lake Fryxell (Lancaster, 2002), indicating that intergranular transport is an important sediment filtering mechanism on ice-covered lakes of the dry valleys. Fine sand around Lake Fryxell contains elevated levels of nitrogen and organic material (Fritsen et al., 2000), and fine sediment in the ice cover of Lake Bonney contains elevated levels of biologically required phosphorus (Priscu et al., 2005), of which the lake plankton are depleted (Priscu, 1995; Dore and Priscu, 2001). Therefore, intergranular sediment transport contributes to the mobility of nutrients in the dry valley lake ice covers.

Our summer 2007–2008 measurements at Lake Fryxell indicate that sediment ablation is a controlling factor of isostatic lake ice adjustment. The ice thickness of 4.6 m and piezometric depth of 0.4 m that we measured in late November (Fig. 7) is consistent with ice in hydrostatic equilibrium. If a fraction α of an ice surface covered in sediment ablates downward a distance h, it will rebound upward a distance $\alpha h \rho_w{}^{-1} \rho_i$ if hydrostatic equilibrium is locally maintained, with a corresponding drop in piezometric level as measured from the ice. This argument assumes that the ice cover density is homogeneous, that the ice surface remains above the piezometric level (which we generally observed), and that meltwater from the ablated ice either evaporates or infiltrates down into the lake water column. This predicted drop in piezometric level is consistent with our observations (assuming $\alpha \sim 0.5$) between November and January (Figs. 1, 6–7). However, the sudden piezometric level drop between 18 and 28 December (Fig. 7) was not synchronous with the rate of surface ablation. Three working hypotheses were considered to explain this. One is a sudden downward flushing of meltwater from shallow ice. For this to be the primary cause, the meltwater must constitute ~50% of the mass above the piezometric level, which seems too high based on our observations. The second hypothesis is an upward flooding of lake water along melting pores in the ice cover. Two findings make this unlikely: drill holes began filling with lake water in mid-November, over a month before the change in piezometric level, and flooding of pores with water would increase the effective density of ice and cause it to sink, not rise. The third, and probably most accurate, hypothesis is as follows. Ice surface ablation by heterogeneously distributed sediment (e.g., Fig. 6d) will lead to lateral variations in ice cover mass, causing internal shear stresses to develop in the ice to balance the buoyancy force from the lake water. This would cause sections of depressed (ablated) ice to rebound upward relative to less depressed ice. Such rebound may occur suddenly, as we observed (Fig. 7), once the shear strength of ice is exceeded. This process may be accompanied by the formation of fractures, which in turn may play a role in sediment "dumping" (Simmons et al., 1986; Squyres et al., 1991). In any case, a sudden shift in piezometric level relative to the ice surface signals either an abrupt change in ice permeability, ice position, or combination thereof. These changes are all directly related to the permeability of ice to sediment and nutrients.

Conclusion

Our laboratory and modeling studies indicate that sediment can melt downward through perennial lake ice in the dry valleys to an equilibrium depth of ~2 m in two melt seasons. This depth is consistent with maximum sediment depths observed in the ice covers (Nedell et al., 1987; Squyres et al., 1991; Priscu et al., 1998; Fritsen and Priscu, 1998), and is consistent with previous findings that fractures and conduits are necessary for delivering sediment to the lake bottom (Simmons et al., 1986; Nedell et al, 1987; Squyres et al., 1991). Results of our model parameter analysis indicate that snow coverage and ablation rates are important controls on the maximum depth of sediment melt.

Our laboratory experiments revealed a ~0.1 g m^{-2} hr^{-1} flux of fine-grained sand along ice grain boundaries, translating to an annual flux of 100 g m^{-2} yr^{-1}. This is within an order of magnitude of average fluxes to lake bottom sediment traps in Lake Hoare (Simmons et al., 1986; Nedell et al., 1987; Squyres et al., 1991), and is greater than the flux of eolian sediment on Lake Fryxell (Lancaster, 2002). These findings indicate that the capacity of the ice to transport mud along grain boundaries is comparable to the annual supply, indicating an important natural sorting mechanism.

Sediment residing on the lake ice loses more heat than subsurface sediment. This heat transfer includes losses to air averaging −5 to −8 °C during summers (Doran et al., 2002b; Fountain et al., 1999) and greater thermal conduction in shallower, colder ice. Surficial sediment therefore melts through ice slower than subsurface sediment. The sediment introduced to Lake Fryxell remained on the ice surface during our field observations; eventually, it will descend below the surface by mechanisms including preferential transport along grain boundaries and the infilling and subsequent freezing of lake water in surface depressions.

Piezometric lake level changes relative to the ice cover provide information about the timing of structural and hydrological changes in the ice. Our analyses of these data indicate that the extent and heterogeneity of surface ablation by sediment is an important control on the timing and configuration of isostatic adjustment of lake ice. These changes are considered to be a reflection of the ice cover's ability to reach a state of local hydrostatic equilibrium.

Acknowledgments

The laboratory work was conducted at Montana State University, Bozeman, and supported by NSF-OPP grants 0085400 and 0346272. The fieldwork was supported by NSF-OPP grants 432595 and 0631494 awarded to Priscu. Thanks to C. Jaraula and A. Hillegas for weighing meltwater sediments, to M. LaRue (image compilation) and B. Herried (map compilation) of the Antarctic Geospatial Information Center for the Lake Fryxell satellite image, and to M. Lizotte for collecting the albedo data. We appreciate the field assistance from M. Dieser, A. Chiuchiolo, and M. Sabacka (Fig. 6c photo), and discussions with K. Welch, M. Hoffman, and C. McKay. This manuscript benefited greatly from the comments by three anonymous reviewers.

References Cited

Adams, E. E., Priscu, J. C., Fritsen, C. H., Smith, S. R., and Brackman, S. L., 1998: Permanent ice covers of the McMurdo Dry Valley lakes, Antarctica: bubble formation and metamorphism. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 281–295.

Barrett, J. E., Virginia, R. A., Wall, D. H., Doran, P. T., Fountain, A. G., Welch, K. A., and Lyons, W. B., 2008: Persistent effects of a discrete warming event on a polar desert ecosystem. *Global Change Biology*, 14(10): 2249–2261.

Brandt, R. E., and Warren, S. G., 1993: Solar-heating rates and temperature profiles in Antarctic snow and ice. *Journal of Glaciology*, 39(131): 99–110.

Bromley, A. M., 1985: Weather observations Wright Valley, Antarctica. New Zealand Meteorological Service, Information Publication 11*:* 37 pp. ISSN 0110-9761.

Bull, C., 1966: Climatological observations in ice-free areas of southern Victoria Land, Antarctica. *In* Rubin, M. J. (ed.), *Studies in Antarctic Meteorology*. Washington, D.C.: American Geophysical Union, 177–194.

Chinn, T. J., 1993: Physical hydrology of the dry valley lakes. *In* Green, W. J., and Friedman, E. I. (eds.), *Physical and Biogeochemical Processes in Antarctic Lakes*. Washington, D.C.: American Geophysical Union, 1–51.

Clow, G. D., McKay, C. P., Simmons, G. M. Jr., and Wharton, R. A. Jr., 1988: Climatological observations and predicted sublimation rates at Lake Hoare, Antarctica. *Journal of Climate*, 1(7): 715–728.

Conovitz, P. A., McKnight, D. M., MacDonald, L. H., Fountain, A. G., and House, H. R., 1998: Hydrologic processes influencing streamflow variation in Fryxell Basin. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 93–108.

Doran, P. T., Wharton, R. A. Jr., Des Marais, D. J., and McKay, C. P., 1998: Antarctic paleolake sediments and the search for extinct life on Mars. *Journal of Geophysical Research–Planets*, 103(E12): 28,481–28,493.

Doran, P. T., Priscu, J. C., Lyons, W. B., Walsh, J. E., Fountain, A. G., McKnight, D. M., Moorhead, D. L., Virginia, R. A., Wall, D. H., Clow, G. D., Fritsen, C. H., McKay, C. P., and Parsons, A. N., 2002a: Antarctic climate cooling and terrestrial ecosystem response. *Nature*, 415(6871): 517–520.

Doran, P. T., McKay, C. P., Clow, G. D., Dana, G. L., Fountain, A. G., Nylen, T., and Lyons, W. B., 2002b: Valley floor climate observations from the McMurdo Dry Valleys, Antarctica. *Journal of Geophysical Research*, 107(D24): 4772, doi:10.1029/2001JD002045.

Dore, J. E., and Priscu, J. C., 2001: Phytoplankton phosphorous deficiency and alkaline phosphatase activity in the McMurdo Dry Valley lakes, Antarctica. *Limnology and Oceanography*, 46(6): 1331–1346.

Drewry, D. J., 1972: Qualitative assessment of dirt-cone dynamics. *Journal of Glaciology*, 11(63): 431–446.

Fetter, C. W., 1999: *Contaminant Hydrogeology.* Second edition. Upper Saddle River, New Jersey: Prentice Hall.

Foreman, C. M., Wolf, C. F., and Priscu, J. C., 2004: Impact of episodic warming events on the physical, chemical and biological relationships of lakes in the McMurdo Dry Valleys, Antarctica. *Aquatic Geochemistry*, 10(3): 239–268.

Fountain, A. G., Lyons, W. B., Burkins, M. B., Dana, G. L., Doran, P. T., Lewis, K. J., McKnight, D. M., Moorhead, D. L., Parsons, A. N., Priscu, J. C., Wall, D. H., Wharton, R. A., and Virginia, R. A., 1999: Physical controls on the Taylor Valley ecosystem, Antarctica. *BioScience*, 49(12): 961–971.

Fountain, A. G., Tranter, M., Nylen, T. H., Lewis, K. J., and Mueller, D. R., 2004: Evolution of cryoconite holes and their contribution to meltwater runoff from glaciers in the McMurdo Dry Valleys, Antarctica. *Annals of Glaciology*, 50(168): 35–45.

Friedmann, E. I., 1982: Endolithic microorganisms in the Antarctic cold desert. *Science*, 215(4536): 1045–1053.

Fritsen, C. H., and Priscu, J. C., 1998: Cyanobacterial assemblages in permanent ice covers on Antarctic lakes: distribution, growth rate, and temperature response of photosynthesis. *Journal of Phycology*, 34(4): 587–597.

Fritsen, C. H., Adams, E. E., McKay, C. P., and Priscu, J. C., 1998: Permanent ice covers of the McMurdo Dry Valleys, Antarctica: liquid water contents. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 269–280.

Fritsen, C. H., Grue, A. M., and Priscu, J. C., 2000: Distribution of organic carbon and nitrogen in surface soils in the McMurdo Dry Valleys, Antarctica. *Polar Biology*, 23: 121–128.

Gribbon, P. W. F., 1979: Cryoconite holes on Sermikavsak, West Greenland. *Journal of Glaciology*, 22(86): 177–181.

Henderson, R. A., Prebble, W. M., Hoare, R. A., Popplewell, K. B., House, D. A., and Wilson, A. T., 1966: An ablation rate for Lake Fryxell, Victoria Land, Antarctica. *Journal of Glaciology*, 6(43): 129–133.

Hendy, C. H., 2000: The role of polar lake ice as a filter for glacial lacustrine sediments. *Geografiska Annaler*, 82A(2–3): 271–274.

Howard-Williams, C., Schwarz, A.-M., Hawes, I., and Priscu, J. C., 1998: Optical properties of the McMurdo Dry Valley lakes, Antarctica. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 189–203.

Jaraula, C. M. B., Kenig, F., Doran, P. T., Priscu, J. C., and Welch, K. A., 2008: SPME–GCMS study of the natural attenuation of aviation diesel spilled on the perennial ice cover of Lake Fryxell, Antarctica. *Science of the Total Environment*, 407(1), doi:10.1016/j.scitotenv.2008.07.064.

Lancaster, N., 2002: Flux of eolian sediment on the McMurdo Dry Valleys, Antarctica: a preliminary assessment. *Arctic, Antarctic, and Alpine Research*, 34(3): 318–323.

Lizotte, M. P., and Priscu, J. C., 1992: Photosynthesis irradiance relationships in phytoplankton from the physically stable water column of a perennially ice-covered lake (Lake Bonney, Antarctica). *Journal of Phycology*, 28(2): 179–185.

Lucchitta, B. K., and Ferguson, H. M., 1983: Chryse Basin channels: low-gradients and ponded flows. *Journal of Geophysical Research*, 88(Supplement): A553–A568.

McIntyre, N. F., 1984: Cryoconite hole thermodynamics. *Canadian Journal of Earth Sciences*, 21(2): 152–156.

McKay, C. P., 1986: Exobiology and future Mars missions: the search for Mars' earliest biosphere. *Advances in Space Research*, 6(12): 269–285.

McKay, C. P., Clow, G. D., Wharton, R. A. Jr., and Squyres, S. W., 1985: Thickness of ice on perennially frozen lakes. *Nature*, 313: 561–562.

McKay, C. P., Clow, G. D., Andersen, D. T., and Wharton, R. A. Jr., 1994: Light transmission and reflection in perennially ice-covered Lake Hoare, Antarctica. *Journal of Geophysical Research*, 99(C10): 20,427–20,444.

McKay, C. P., Andersen, D. T., Pollard, W. H., Heldmann, J. L., Doran, P. T., Fritsen, C. H., and Priscu, J. C., 2005: Polar lakes, streams, and springs as analogs for the hydrological cycle on Mars. *In* Tokano, T. (ed.), *Water on Mars and Life. Advances in Astrobiology and Biogeophysics*. Berlin/Heidelberg: Springer-Verlag, 219–233.

McKnight, D. M., Alger, A., Tate, C. M., Shupe, G., and Spaulding, S., 1998: Longitudinal patterns in algal abundance and species distribution in meltwater streams in Taylor Valley, southern Victoria Land, Antarctica. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 109–127.

MCM LTER: McMurdo Dry Valleys Long Term Ecological Research data sets. http://www.mcmlter.org/data_home.htm, last accessed 15 December 2008.

Morgan-Kiss, R. M., Priscu, J. C., Pocock, T., Gudynaite-Savitch, L., and Huner, N. P. A., 2006: Adaptation and acclimation of photosynthetic microorganisms to permanently cold environments. *Microbial and Molecular Biology Reviews*, 70: 222–252.

Nedell, S. S., Andersen, D. W., Squyres, S. W., and Love, F. G., 1987: Sedimentation in ice-covered Lake Hoare, Antarctica. *Sedimentology*, 34(6): 1093–1106.

Paerl, H. W., and Priscu, J. C., 1998: Microbial phototrophic, heterotrophic and diazotrophic activities associated with aggregates in the permanent ice cover of Lake Bonney, Antarctica. *Microbial Ecology*, 36(3): 221–230.

Parker, B. C., Simmons, G. M. Jr., Love, F. G., Wharton, R. A. Jr., and Seaburg, K. G., 1981: Modern stromatolites in Antarctic dry valley lakes. *BioScience*, 31(9): 656–661.

Parker, B. C., Simmons, G. M. Jr., Seaburg, K. G., Cathey, D. D., and Allnut, F. C. T., 1982: Comparative ecology of plankton communities in seven Antarctic oasis lakes. *Journal of Plankton Research*, 4(2): 271–286.

Preston, D. W., and Dietz, E. R., 1991: *The Art of Experimental Physics*. Toronto, Canada: John Wiley & Sons, 432 pp.

Priscu, J. C., 1995: Phytoplankton nutrient deficiency in lakes of the McMurdo Dry Valleys, Antarctica. *Freshwater Biology*, 34(2): 215–227.

Priscu, J. C. (ed.), 1998: *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union.

Priscu, J. C., Fritsen, C. H., Adams, E. E., Giovannoni, S. J., Paerl, H. W., McKay, C. P., Doran, P. T., Gordon, D. A., Lanoil, B. D., and Pinckney, J. L., 1998: Perennial Antarctic lake ice: an oasis for life in a polar desert. *Science*, 280: 2095–2098.

Priscu, J. C., Adams, E. E., Paerl, H. W., Fritsen, C. H., Dore, J. E., Lisle, J. T., Wolf, C. F., and Mikucki, J. A., 2005: Perennial Antarctic lake ice: a refuge for cyanobacteria in an extreme environment. *In* Castello, J. D., and Rogers, S. O. (eds.), *Life in Ancient Ice*. Princeton, New Jersey: Princeton University Press, 22–49.

Priscu, J. C., Christner, B. C., Dore, J. E., Westley, M. B., Popp, B. N., Casciotti, K. L., and Lyons, W. B., 2008: Extremely supersaturated N_2O in a perennially ice-covered Antarctic lake: molecular and stable isotopic evidence for a biogeochemical relict. *Limnology and Oceanography*, 53: 2439–2450.

Schlatter, T. W., 1972: The local surface energy balance and subsurface temperature regime in Antarctica. *Journal of Applied Meteorology*, 11(7): 1048–1062.

Sharp, R. P., 1949: Studies of superglacial debris on valley glaciers. *American Journal of Science*, 247(5): 289–315.

Simmons, G. M. Jr., Wharton, R. A. Jr., McKay, C. P., Nedell, S., and Clow, G., 1986: Sand/ice interactions and sediment deposition in perennially ice-covered Antarctic lakes. *Antarctic Journal of the United States*, 21(5): 217–220.

Simmons, G. M. Jr., McKay, C. P., and Wharton, R. A. Jr., 1987: Ice thickness changes on Lake Hoare, southern Victoria Land, Antarctica. *Antarctic Journal of the United States*, 22(5): 235–236.

Spigel, R. H., and Priscu, J. C., 1998: Physical limnology of the McMurdo Dry Valley lakes. *In* Priscu, J. C. (ed.), *Ecosystem Dynamics in a Polar Desert, the McMurdo Dry Valleys, Antarctica*. Washington, D.C.: American Geophysical Union, 153–187.

Squyres, S. W., 1989: Urey prize lecture: water on Mars. *Icarus*, 79(2): 229–288.

Squyres, S. W., Andersen, D. W., Nedell, S. S., and Wharton, R. A. Jr., 1991: Lake Hoare, Antarctica: sedimentation through a thick perennial ice cover. *Sedimentology*, 38(2): 363–379.

Thompson, D. C., Craig, R. M. F., and Bromley, A. M., 1971: Climate and surface heat balance in an Antarctic dry valley. *New Zealand Journal of Science*, 14(2): 245–251.

Ugolini, F. C., 1970: Antarctic soils and their ecology. *In* Holdgate, M. W. (ed.), *Antarctic Ecology,* Vol. 2. London, U.K.: Academic Press, 673–692.

Wagner, B., Melles, M., Doran, P. T., Kenig, F., Forman, S. L., Pierau, R., and Allen, P., 2006: Glacial and postglacial sedimentation in the Fryxell basin, Taylor Valley, Southern Victoria Land, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 241(2): 320–337.

Wharton, R. A. Jr., Parker, B. C., Simmons, G. M. Jr., Seaburg, K. G., and Love, F. G., 1982: Biogenic calcite structures forming in Lake Fryxell, Antarctica. *Nature*, 295(5848): 403–405.

Wharton, R. A. Jr., McKay, C. P., Simmons, G. M. Jr., and Parker, B. C., 1985: Cryoconite holes on glaciers. *BioScience*, 35(8): 499–503.

Wharton, R. A. Jr., McKay, C. P., Clow, G. D., Andersen, D. T., Simmons, G. M. Jr., and Love, F. G., 1992: Changes in ice cover thickness and lake level of Lake Hoare, Antarctica: implications for local climate change. *Journal of Geophysical Research*, 97(C3): 3503–3513.

Wharton, R. A. Jr., McKay, C. P., Clow, G. D., and Andersen, D. T., 1993: Perennial ice covers and their influence on Antarctic lake ecosystems. *In* Green, W. J., and Friedman, E. I. (eds.), *Physical and Biogeochemical Processes in Antarctic Lakes*. Washington, D.C.: American Geophysical Union, 53–70.

Wilson, A. T., 1981: A review of the geochemistry and lake physics of the Antarctic dry areas. *In* McGinnis, L. D. (ed.), *Dry Valley Drilling Project*. Washington, D.C.: American Geophysical Union, 185–192.

Wiscombe, W. J., and Warren, S. G., 1980: A model for the spectral albedo of snow. I: Pure snow. *Journal of the Atmospheric Sciences*, 37(12): 2712–2733.

MS accepted July 2009