Vapor transport and sublimation on Mullins Glacier, Antarctica

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Abstract

We utilize an environmental chamber capable of recreating the extreme polar conditions of the McMurdo Dry Valleys (MDV) of Antarctica to investigate the sublimation rate of the Mullins Valley debris-covered glacier (hereafter Mullins Glacier), reportedly one of the oldest debris-covered alpine glaciers in the world. We measure ice loss via sublimation beneath sediment thicknesses ranging from 0 to 69 mm; from this, we determine an effective diffusivity for Fickian vapor transport through Mullins till of $(5.2 \pm 0.5) \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ at -10°C . We use this value, coupled with micrometeorological data from Mullins Valley (atmospheric temperature, relative humidity, and soil temperature) to model the sublimation rate of buried glacier ice near the terminus of Mullins Glacier, where the overlying till thickness approaches 70 cm. We find that the ice-lowering rate during the modeled year (2011-2012) was 0.066 mm under 70 cm of till, a value which is in line with previous estimates for exceedingly slow rates of ice sublimation. These results provide further evidence supporting the probable antiquity of Mullins Glacier ice and overall landscape stability in upland regions of the MDV.

Keywords: Sublimation, buried ice, McMurdo Dry Valleys, Antarctica, diffusion

1. Introduction

The McMurdo Dry Valleys (MDV) of Antarctica have remained predominately ice-free and under hyper-arid polar desert climate conditions since the middle Miocene (Denton et al., 1993; Marchant et al., 1993; Marchant et al., 1996; Marchant et al., 2002; Sugden et al., 1999; Valletta et al., 2015). The preservation of *in situ* volcanic ash deposits > 10.0 Ma in the Quartermain Mountains (Marchant et al., 1993; Sugden et al., 1995a), as well as the concentration of meteoric beryllium-10 in ancient sediments (Valletta et al., 2015), demonstrate that saturated active-layer cryoturbation - typical of most periglacial regions in the world - has been essentially absent from much of the western MDV over this interval. These findings corroborate the slow rates of landscape change and long-term stability of the region that is afforded by cosmogenic nuclide analyses of surface rocks (Schäfer et al., 2000; Schäfer et al., 1999; Margerison et al., 2005) and glacial-geologic mapping (Lewis and Ashworth, 2015). Moreover, the presence of *in-situ* ashfall in supraglacial debris suggests that, in places, buried glacier ice has survived for ~8.1 Ma (Sugden et al., 1995a). If such ancient ice could be analyzed reliably for gas content it would greatly extend our climate record beyond that possible through the analysis of existing ice cores (e.g., EPICA Dome C, Vostok, GISP). This possibility, along with recent field data suggesting that numerous buried glaciers survive in the coldest and driest parts of the Dry Valleys / Transantarctic Mountains (Mackay and Marchant, 2016) has prompted renewed interest in defining the precise conditions that foster long-term ice preservation in hyper-arid deserts (Kowalewski et al., 2006; Yau et al., 2015; Mackay et al., 2014). The results have been mixed, with modeled estimates for sublimation rates of the same buried ice masses spanning two orders of magnitude, from 10^{-3} m yr⁻¹ (Hindmarsh et al., 1998) to 10^{-5} m yr⁻¹ (Kowalewski et al., 2012; Liu et al., 2015) (see also Hagedorn et al., 2007; Hindmarsh et al., 1998; Kowalewski et al., 2012; Kowalewski et al., 2006; Kowalewski et al., 2011; Schorghofer,

2005). If the former rates are correct, then it is unlikely that buried glaciers could survive for $>\sim 1$ Ma; if the latter are correct, however, then pockets of buried ice could be preserved for several million years. As yet, there have been no attempts to address this problem through experimental analyses.

In this study, we examine the sublimation rate of Mullins Glacier, a debris-covered alpine glacier located in the inland portion of the MDV, using an experimental approach that measures ice loss via sublimation in an environmental chamber. In its central region, Mullins Glacier is > 1.6 Ma (Yau et al., 2015; Mackay and Marchant, 2016); near its terminus, it may be > 3 Ma (Mackay and Marchant, 2016).

2. Setting

Mullins Glacier is covered with a thin supraglacial till (Mullins till) that is derived from rockfall at the valley head; the till and underlying glacier ice move slowly down valley, decelerating from a maximum velocity of ~ 40 mm yr⁻¹ near the valley head to < 1 to 2 mm yr⁻¹ on the floor of upper Beacon Valley (Fig. 1) (essentially stagnant, e.g., within measurement error of signals from interferometric synthetic aperture radar; Rignot et al., (2002)). Mullins till is composed primarily of Ferrar Dolerite ($>93\%$) and undifferentiated sandstone fragments (Kowalewski et al., 2011). The mean annual and summertime atmospheric temperatures at the study site on the glacier are -23°C and -11°C respectively (Kowalewski et al., 2011; Mackay et al., 2014), and snowfall is < 50 mm yr⁻¹ water equivalent (Fountain et al., 2010). Strong katabatic winds flow from the East Antarctic Ice Sheet (EAIS) at speeds commonly approaching 50 km hr⁻¹ (Nylén et al., 2004; Speirs et al., 2010).

Figure 1. Location of Mullins Glacier. Mullins Glacier occupies the region within the white box, in the lower left portion of the figure; the brown areas outlined in yellow highlight the stable upland zone of the McMurdo Dry Valleys, e.g., the region with negligible saturated active-layer cryoturbation (Marchant and Head, 2007). Inset map shows location of the McMurdo Dry Valleys (MDV); see Figure 2 for details. Figure reproduced from Lamp et al. (2017).

The ice in the upper reaches of Mullins Glacier (within ~ 4 km from the valley headwall) is relatively clean, showing only dispersed fines and episodic layers of concentrated debris; beyond ~4 km, the englacial content in Mullins Glacier increases markedly, with visual estimates of shallow ice cores and buried-ice surfaces in soil excavations, along with interpretations of ground penetrating radar, indicating up to 50% gravel and cobble sized clasts by volume (Mackay et al., 2014). The thickness of Mullins till generally shows a corresponding increase with increasing distance from the headwall (Mackay et al., 2014); at its maximum, Mullins till reaches ~70 cm in thickness in central and upper Beacon Valley (Kowalewski et al., 2011). Throughout its length, Mullins till is dotted with sublimation polygons, each of which increases in size and maturity with distance down glacier (Marchant et al., 2002). Previous studies have shown that these polygons play an important role in modulating subsurface ice loss (Kowalewski et al., 2012; Marchant et al., 2002): initially, sublimation rates at polygon troughs are high, due

in part to elevated ice loss within unprotected thermal contraction cracks at these sites; as troughs deepen, however, they are in shadow most of the day and experience relatively cool temperatures. Further, deep troughs become preferred sites for snow entrapment, elevating local relative humidity (RH) (Marchant et al., 2002; Kowalewski et al., 2012). The relatively high RH values and cooler atmospheric temperatures tend to retard underlying ice sublimation (Kowalewski et al., 2006; Schorghofer, 2005). The result is that long-term sublimation rates for buried glacier ice with overlying sublimation polygons is limited by the rate of ice loss at polygon centers (Kowalewski et al., 2012).

Figure 2. Mullins Glacier, viewed as a hill-shaded relief map from airborne LiDAR (2-m resolution (Schenk et al., 2004)) draped on top of multispectral satellite imagery (GeoEye01, 1.5-m resolution, January 2004) in headwall region (lower left corner) (adapted from Mackay et al., 2014). Flow is from left (Mullins Valley) to right (Beacon Valley). Samples at location JKS-10-020 were collected from the middle of a polygon ~1 km from the diffuse terminus of Mullins Glacier, central Beacon Valley. Data collected at the long-duration micrometeorological station (Met. Station) is used as input for numerical models described in Section 6. The depth of buried glacier ice at the Met station is ~25 cm, whereas at JKS-10-20 it is ~70 cm.

3. Theoretical background

We assume that Fickian diffusion, characterized by impacts between molecules and driven by concentration gradients, is the dominate vapor transport mechanism from buried ice to the atmosphere at polygon centers in Mullins till (Schorghofer and Aharonson, 2005). Fickian processes dominate diffusive vapor transport through porous media when pore sizes are greater than the mean free path of vapor molecules (Ho and Webb, 2006). Previous studies of sediment in the MDV have shown that Knudsen diffusion, which takes into account molecular interactions with pore walls, can be neglected (Hagedorn et al., 2007; McKay et al., 1998). Thus, for Mullins till, Knudsen diffusion is insignificant. Additionally, we do not include the effects of thermally-driven vapor diffusion or advection, which previous researchers have shown to be less important than Fickian diffusion on annual (and longer) timescales (Schorghofer, 2005).

Fick's first law of diffusion is given as:

$$J = -D_f \frac{\partial \rho_{vap}}{\partial z} \quad (\text{Eq. 1})$$

where J is the diffusive flux ($\text{mol m}^{-2} \text{s}^{-1}$), D_f is the Fickian diffusion coefficient ($\text{m}^2 \text{s}^{-1}$), ρ_{vap} is water vapor density (mol m^{-3}) and z is length (m). For diffusion through porous media, this is rewritten as:

$$J = -D_{eff} \frac{\partial \rho_{vap}}{\partial z} \quad (\text{Eq. 2})$$

where D_{eff} is the effective diffusion coefficient ($\text{m}^2 \text{s}^{-1}$), and,

$$D_{eff} = \frac{\phi D_f}{\tau} \quad (\text{Eq. 3})$$

where ϕ is the porosity of the sediment, and τ is the tortuosity. Assuming a linear decrease in ρ_{vap} throughout the sediment column, and saturated vapor density at the ice-sediment interface, Eq. 2 can be rewritten as:

$$J = -D_{eff} \frac{\rho_{sat} - \rho_{atm}}{L} \quad (\text{Eq. 4})$$

where L is the sediment thickness, ρ_{sat} is the saturated vapor density at the temperature of the ice-sediment interface, and ρ_{atm} is the vapor density in the atmosphere.

We can relate vapor density (ρ_{sat} , ρ_{atm}) to vapor pressure (P_{sat} , P_{atm}) using the ideal gas law:

$$\rho_{vap} = \frac{P_{vap} M_w}{R T} \quad (\text{Eq. 5})$$

where P_{vap} is vapor pressure, M_w is the molecular mass of water (18.02 g mol⁻¹), R is the gas constant (8.314 J mol⁻¹ K⁻¹), and T is temperature.

Substituting Eq. 5 into Eq. 4 we find:

$$J = -\frac{D_{eff} M_w}{L R} \left[\frac{P_{sat}}{T_{ice}} - \frac{P_{atm}}{T_{atm}} \right] \quad (\text{Eq. 6})$$

where P_{sat} is the saturation vapor pressure at the temperature of the ice-sediment interface (T_{ice}), P_{atm} is the vapor pressure in the atmosphere and T_{atm} is the atmospheric temperature. The value of P_{sat} over ice can be approximated by the polynomial equation given in Lowe (1977) and Hagedorn et al. (2007):

$$P_{sat} = (A + T(B + T(C + T(D + T(E + T(F + GT)))/H) \quad (\text{Eq. 7})$$

where P_{sat} (Pa) is the saturated vapor pressure at T (°C) over ice. The coefficient values are: A = 6.109177956, B = 0.503469897, C = 1.886013408 x 10⁻², D = 4.176223716 x 10⁻⁴, E = 5.824720280 x 10⁻⁶, F = 4.838803174 x 10⁻⁸, G = 1.838826904 x 10⁻¹⁰, and H = 0.009998.

We can now divide J (Eq. 6) by the density of ice (ρ_{ice}) to convert it from a flux to a sublimation rate, E , in units of length per time. Rearranging, this gives:

$$E = -\frac{D_{eff} M_w P_{sat}}{L R T_{ice} \rho_{ice}} \left[1 - \frac{T_{ice} P_{atm}}{T_{atm} P_{sat}} \right] \quad (\text{Eq. 8})$$

Eq. 8 is used in conjunction with our experimental measurements of sublimation loss (see below) to estimate the effective diffusion coefficient in Mullins till and to predict the annual loss of ice from Mullins Glacier using field measured values of temperature, RH, and air pressure.

4. Experimental methods

4.1 Physical properties of Mullins till

4.1.1 Generation of the Mullins Composite (MC) standard

MC consists of a conglomeration of Mullins till sourced from separate sample sites on Mullins Glacier; the composite was sorted by grain size, and then recombined to match a representative sediment sample with a prescribed grain size distribution. The representative sample for MC is JKS-10-020, which comes from near the glacier terminus (Figure 2) at ~ 1250 m.a.s.l. and best represents the till texture at the location on Mullins Glacier where some of the oldest ice is preserved (Kowalewski et al., 2011). JKS-10-020 was taken from within 10 cm of the till surface. Although till texture varies with depth to the buried-ice interface, distance from the headwall, and location with respect to polygonal surfaces, we believe this sample provides an adequate representation of Mullins till for the purposes of this initial study.

To prepare the MC, the grain size distribution of our representative sample, JKS-10-020, was first determined by sieving samples in the field (to separate the < 16 mm fraction from the > 16 mm fraction) and then sieving the sample (dry) at Boston University through a stack of 9 sieves: 8 mm, 4 mm, 2 mm, 1 mm, 500 μm , 250 μm , 125 μm , 63 μm , and pan (< 63 μm). From this analysis, we determined that JKS-10-020 is a coarse-grained, well-graded sand-and-gravel mixture containing < 2% silt and clay sized particles (Figure 3, see also table provided in supplementary material). This grain size distribution was replicated in MC by combining

additional Mullins till samples, sieving as described above, and then combining the fractions to match the weight percentages in JKS-10-020. A new MC sample was prepared prior to each sublimation experiment in order to ensure that the grain size distribution for each experiment was constant and unaffected by gravitational settling during storage. The Mullins till samples were stored in Whirl-Pak[®] bags and sealed plastic containers from the time of their initial removal from the field site until their use in this study which minimized the duration of exposure to the atmosphere and the amount of water adsorption/desorption.

Only particle sizes < 8 mm were used in the experiments; this maximum grain-size value was chosen due to the relatively narrow inner diameter (3.25 inches or 82.55 mm) of the sample containers used in our sublimation experiments (Figure 4). Kowalewski et al. (2011) directly measured the *in situ* porosity of Mullins till, reporting values ranging from 0.29-0.31. To determine porosity in this study, we measured the bulk density (ρ_B) of the MC sediment (~1896 kg m⁻³); assuming a grain density (ρ_G) of ~ 2700 kg m⁻³, the porosity (ϕ) of MC is ~29.8% ($\phi = 1 - \rho_B/\rho_G$), which is in accord with porosity at polygon centers reported Kowalewski et al. (2011), e.g., ~ 29.5%.

Figure 3. Cumulative particle size distribution of JKS-10-20 and the Mullins Composite (MC).

4.1.2 Salts in Mullins till/MC

A 50 g sample of MC sediment was rinsed with deionized water (100 mL), and the supernatant allowed to evaporate, leading to the precipitation of salt crystals. These crystals were analyzed in Boston University's X-ray Diffraction Lab on a Rigaku MiniFlex Powder Diffractometer. The diffraction patterns, analyzed using the Jade 9 data analysis program, showed evidence of nitratine (NaNO_3), gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$), and halite (NaCl) (see Figure S1 in the supplementary material). Because of the manner in which we admixed samples to create MC, the distribution of these salts in MC is most likely uniform with depth; however, discrete salt layers are common in sublimation till in the western MDV (Bao and Marchant, 2006; Bao et al., 2008). The inclusion of salt horizons in our experiments would likely result in lower sublimation rates as these low-porosity layers hinder vapor transport in the till column (Kowalewski et al., 2012).

4.2 Environmental chamber setup

We used an ESPEC environmental chamber, model EPX-3H, coupled with an oil-free Atlas Copco air compressor and dry-air purge system (capable of producing air with a dew point $< -30^\circ\text{C}$) to analyze sublimation and vapor transport through the MC standard. The internal chamber compartment (Figure 4B) is equipped with two analytical balances set on anti-vibration pads, a suite of Onset Computer Corporation HOBO micrometeorological sensors (air temperature, relative humidity (RH) and air pressure; see specifications in supplementary material), and a Plexiglas air shield to block air flow from the chamber's air inlet from passing directly over the sample. The ice core/sediment containers are ~ 30 cm tall and made from rigid, clear butyrate tubing with a 3.25 in (82.55 mm) inner diameter and capped at one end (Figure 4C). Readings from an anemometer mounted in the chamber during experiments showed that the

air speed at the height of the samples was generally between 0-0.1 m/s. The chamber temperature was set to -10°C, varying $\leq 0.1^\circ\text{C}$ during the course of each experiment. At -10°C, RH values varied between 12%-22% over the course of all experiments, reflecting fluctuations in the ambient air conditions at the compressor intake; however, the RH during each individual experiment remained relatively stable (see Supplemental Information).

Figure 4. Experimental setup of sublimation experiments. (A) The ESPEC environmental Chamber, model EPX-3H. (B) The inside of the test chamber. The micrometeorological sensors and balance connections are fed through a port on the left-hand side of the chamber which is insulated with foam. A balance is seated on an anti-vibration pad, and is surrounded with a Plexiglas air shield; a second air shield, which provides additional protection, is not pictured. A battery powered anemometer hangs above the balance. (C) A typical butyrate sample container (~30 cm tall) that would hold a core of ice and overlying Mullins Composite of varying thickness.

4.3 Balances

Two Mettler Toledo MS4002S laboratory balances (4200 g capacity and a readability of 0.01 g) were used within the environmental chamber to record changes in the mass of studied ice samples. Prior to conducting individual experiments, we tested the stability of the balances at subzero temperatures. The balances were placed into the chamber at -10 °C for 24 hours, loaded with standard weights of 500 g to 1500 g, and masses were recorded every 60 seconds. The results of this test showed the balances maintained high precision but required a start-up period of ~2-3 hours to enable the scales to equilibrate to the chamber temperatures (Figure S2 in the

supplementary material). Therefore, we exclude the first 3 hours of data in all sublimation experiments in order to account for scale equilibration.

4.4 Experimental procedures

For each experiment, butyrate sample containers (~30 cm tall; Figure 4C) were filled with deionized water to a pre-determined height and frozen at -10°C to produce a ~20-30 cm ice core; if necessary the ice surface was shaved down and a thin layer of water frozen on top to create a flat interface between ice and sediment. MC was then added to the ice surface until the sediment was flush with the rim of the container. To prevent melting at the ice-sediment boundary, the MC was pre-cooled to -10°C before being added to the ice core. Measurements of mass loss were recorded every 60 seconds, and experiments were run for varying lengths of time, ranging from ~24 hours for bare ice to ~2 weeks for the thickest sediment cover. All experiments were run at -10°C in order to (1) be sufficiently far from the melting point of ice, and (2) preserve the life of the analytical balances. Additionally, a MC sample was run by itself in the chamber at -10°C to determine whether desorption from the sediment was in part responsible for any mass loss seen in the sublimation experiments.

At the conclusion of each experiment, the sublimation rate was determined by fitting a linear trend line to the measured mass versus time data. The RH and air pressure for each experiment were determined by averaging the measured RH and air pressure over the portion of the experiment used in the sublimation rate calculation.

6. Results

6.1 Experimental results

Experimental results show that sublimation rates decrease nonlinearly with increased thicknesses of overlying MC samples (Table 1). Mass loss versus time plots for each run are provided in the supplementary material; the data show that our 60 second data-recording interval is sufficient to produce a strong trend, despite chamber vibrations adding noise to the balance readings. We found no evidence for mass loss due to desorption from the isolated MC sediment (no ice) experiment.

6.2 Diffusivity of Mullins till

We use the experimental results and Eq. 8 to determine the effective diffusivity in Mullins till. Assuming steady-state conditions and a constant temperature profile in the sediment during our experiments (i.e., $T_{ice} = T_{atm} = T$), Eq. 8 reduces to:

$$E = -\frac{D_{eff} M_w P_{sat}}{L R T \rho_{ice}} [1 - RH] \quad (\text{Eq. 9})$$

where RH is the relative humidity, equal to P_{atm}/P_{sat} , and ρ_{ice} is the density of ice at -10°C which is $\sim 918.2 \text{ kg m}^{-3}$ (Haynes, 2015). In order to accurately determine the RH at -10°C , the RH values from the Onset HOB0 sensors used in the chamber (which are reported with respect to liquid water) were multiplied by the ratio of the saturated vapor pressure over water to the saturated vapor pressure over ice using the polynomial equations in Lowe (1977).

We correct for fluctuations in RH and air pressure (P_{air}) during our experiments using:

$$E_{corr} = E_{exp} \left[\left(\frac{1 - RH_{corr}}{1 - RH_{exp}} \right) \left(\frac{P_{air,exp}}{P_{air,corr}} \right) \right] \quad (\text{Eq. 10})$$

where E_{corr} is the corrected sublimation rate and E_{exp} is the experimentally determined sublimation rate. Using Eq. 10, each experimentally derived sublimation estimate made at conditions of RH_{exp} , $P_{air,exp}$ can be converted to the expected sublimation rate at consistent RH , P_{air} values. In order to accurately compare the sublimation results between experimental runs, we

corrected the sublimation rate for each experiment using Eq. 10 to represent the following ambient conditions: $T = -10^{\circ}\text{C}$, $RH_{corr} = 40\%$, $P_{air,corr} = 850$ mbar (Table 1).

We use values of E_{corr} at each sediment thickness to approximate the value of D_{eff} for Mullins till according to Eq. 9. We first define a parameter β :

$$\beta = \frac{M_w P_{sat}}{L R T \rho_{ice}} [1 - RH] \quad (\text{Eq. 11})$$

So that

$$E_{corr} = \beta D_{eff} \quad (\text{Eq. 12})$$

From the slope of a E_{corr} versus β plotted for each sediment thickness (Figure 5) we can now estimate the value of D_{eff} for Mullins till. For a chamber temperature of -10°C , 40% RH, and an atmospheric pressure of 850 mbar, our experimental results indicate an effective diffusivity of $(5.2 \pm 0.5) \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ for Mullins till. Next, we estimate the tortuosity - the ratio of the average "tortuous" path length traveled by a molecule through interconnected pores to the apparent straight-line path length (Hillel, 1998) - in the sediment using Eq. 3 with our known value of porosity (0.298, from Section 4.1.1) and an estimate of D_f (in $\text{m}^2 \text{ s}^{-1}$) for water vapor in air given as (Schwartz and Brow, 1951):

$$D_f = (1.87 \times 10^{-10}) T^{2.072} \frac{P_0}{P} \quad (\text{Eq. 13})$$

where $P_0 = 1013$ mbar, and T and P are the values used in our experimental determination of D_{eff} (-10°C and 850 mbar). With our estimated D_{eff} value, we arrive at an estimated tortuosity in the MC sediment of ~ 1.33 .

Table 1. Results from sublimation experiments.

Sediment depth (mm)	Average RH (%)	Average pressure (mbar)	Experimental sublimation rate ($10^{-10} \text{ m s}^{-1}$)	Corrected sublimation rate ^a ($10^{-10} \text{ m s}^{-1}$)	β (10^{-5} m^{-1})
0	21.59	1026	208	192	--
0	21.59	1026	163	150	--
6	20.20	1009	11.8	10.6	23.3
7	12.24	1018	12.4	10.2	19.9

11.5	12.24	1018	9.46	7.75	12.1
12.5	20.20	1009	7.55	6.74	11.1
18	17.31	1028	5.45	4.78	7.76
30.5	17.31	1028	3.50	3.07	4.58
42	17.39	1017	1.91	1.66	3.32
69	17.39	1017	1.10	0.958	2.02

^a Corrected to 40% RH and 850 mbar

Figure 5. Sublimation rate versus β . The value of D_{eff} is approximated by the slope of the line to be $(5.2 \pm 0.5) \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. The R^2 value for the line is 0.93.

6.3 Sublimation rates on Mullins Glacier

We now utilize our estimate of D_{eff} , in conjunction with measured micrometeorological values in Beacon Valley, to estimate the sublimation rate near the terminus of Mullins Glacier in Beacon Valley, where the till thickness approaches 70 cm (Kowalewski et al., 2011). Using an existing multi-year record of air temperature, RH, and pressure, as well as subsurface soil temperatures at depths of 1 to 25 cm in Mullins till, we first estimate the thermal diffusivity of Mullins till and then use this value to calculate the expected till temperature at depths of 40, 50, 60, and 70 cm. We then estimate the average value of D_{eff} , which itself changes with temperature and air pressure, in the sediment column at each time interval. Lastly, we use Eq. 9 to determine the sublimation rate of Mullins Glacier ice at each time step, and then calculate the annual sublimation rate for till thicknesses from 1 to 70 cm.

6.3.1 Micrometeorological data

Multi-year records of air temperature, RH, and pressure at 2 cm above the till surface, as well as subsurface soil temperatures in Mullins till at depths of 1, 10, 20, and 25 cm depth (the local depth to buried glacier ice at the monitoring location) are available from a micrometeorological station located halfway between the glacier headwall and terminus (Figure 2). The data were recorded every 30 minutes using Onset HOBO dataloggers and smart sensors. In this study, we use a year of continuous data from January 1, 2011-January 1, 2012. We select this dataset because we have a continuous, unbroken record for the entire year, and the average values are similar to the multi-year averages in this location (Table 2).

The time series and average micrometeorological data for the year of interest are provided in Figures 6, 7, and Table 2. The RH values for air temperatures $< 0^{\circ}\text{C}$ were corrected for RH over ice as described in section 6.2.

Figure 6. 2011 air pressure and adjusted RH from Mullins Glacier. Time 0 corresponds to January 1, 2011. (a): RH of air 2 cm above the till surface; all subzero RH values are adjusted to report the RH with respect to ice (see text). (b): Air pressure at 2cm above the till surface in mbar.

Figure 7. 2011 air and till temperatures measured on Mullins Glacier. Data series starts at January 1, 2011 and ends on December 31, 2012.

Table 2. Summary of meteorological measurements on Mullins Glacier.

	Temperatures (°C)									RH ¹ (%)	Pressure (mbar)
	Air	1 cm	10 cm	20 cm	25 cm	40 cm	50cm	60 cm	70 cm		
2011 Average	-20.5	-20.2	-20.2	-20.1	-20.4	-20.2	-20.2	-20.2	-20.2	70.2	820
2011 Maximum	2.46	13.1	4.48	-3.07	-5.7	-12.4	-13.5	-14.4	-15.2	101 ¹	854
2011 Minimum	-43.2	-44.9	-40.1	-33.8	-31.9	-28.0	-27.0	-26.1	-25.3	6.75	793
Multi-year average ²	-21.5	-21.2	-21.2	-21.0	-21.3	--	--	--	--	--	--

¹ Corrected (see text); the correction leads to the RH over ice becoming >100 for < 0.8% of the measurements.

² Air temperature data from 12/2008 - 12/2012; Till temperatures from 12/2006 - 12/2012

6.3.2 Estimate of temperatures at depth

In order to approximate the sublimation rate of ice buried under 40 - 70 cm of till, we first approximate the temperature wave at depth using the measured till temperatures (see supplementary material). We find a best fit cosine curve for each temperature wave and then approximate the thermal diffusivity (κ) in the till using:

$$\ln A_z = \ln A_0 - \frac{z}{\sqrt{T\psi\kappa/\pi}} \quad (\text{Eq. 14})$$

where ψ is the period, z is the depth, A_0 is the amplitude of the temperature wave at the surface, and A_z is the amplitude at depth z . We approximate κ by finding the best-fit line to a plot of $\ln(A_z)$ vs. z , and solving for κ from the slope (see supplementary material). This results in an

estimated thermal diffusivity for Mullins till of $\sim 2.1 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$, which is in line with previous estimates for supraglacial till and other sediment in the MDV (Campbell and Claridge, 2006; Kowalewski et al., 2011). We then approximate the temperature wave at $z = 40, 50, 60$, and 70 cm by estimating for each depth the amplitude via Eq. 14, and the phase shift by the best fit slope of the phase shift vs. z from the measured temperatures at shallower depths.

6.3.3 Sublimation model

We now use the measured ($1, 10, 20$, and 25 cm depth) and estimated ($40, 50, 60$, and 70 cm depth) Mullins till temperatures, coupled with the estimate of D_{eff} from the chamber experiments, to approximate the sublimation of Mullins Glacier ice under various thickness of till.

The diffusion coefficient is not constant, but has been found to correlate approximately with temperature and pressure as (Chen and Othmer, 1962):

$$D_f \propto \frac{T^{\frac{3}{2}}}{P} \quad (\text{Eq. 15})$$

We estimate D_{eff} at each depth and time using:

$$D_{eff} = D_{eff,exp} \frac{P_{exp}}{P} \frac{T^{\frac{3}{2}}}{T_{exp}^{\frac{3}{2}}} \quad (\text{Eq. 16})$$

where P_{exp} and T_{exp} are the pressure and temperature at which the experimentally determined value of $D_{eff,exp}$ was calculated, P and T are the average values of air pressure and temperature between each temperature record location. At each 30-minute time interval, we calculate D_{eff} for depth segments of the till column: $0\text{-}1$ cm depth, $1\text{-}10$ cm depth, $10\text{-}20$ cm depth, $20\text{-}25$ cm depth, and $25\text{-}base$ depth, depending on the total thickness of the column modeled. We determine a depth-averaged value of diffusivity ($D_{eff,avg}$) from these values and apply it to the

total sediment column at each time step. The air pressure is assumed to be constant through the till column at each time step.

The density of ice (ρ_{ice}) also varies slightly with temperature. We approximate this variability over the range of 0°C to -50°C using a linear fit to density measurements (Haynes, 2015) resulting in:

$$\rho_{ice} = -0.13514 T_{ice} + 953.719 \quad (\text{Eq. 17})$$

where T_{ice} is the temperature in K at the ice-sediment boundary.

Finally, we solve for the sublimation rate at each 30-minute time step using:

$$E(t) = -\frac{D_{eff,avg}(t) M_w}{z R \rho_{ice}(t)} \left[\frac{P_{sat,z}(t)}{T_z(t)} - \frac{P_{sat,air}(t) RH_{air}(t)}{T_{air}(t)} \right] \quad (\text{Eq. 18})$$

where $P_{sat,z}$, and T_z are the conditions at the base of the till column (the ice table depth) assuming saturated conditions at the ice-sediment interface, and $P_{sat,air}$, RH_{air} , and T_{air} are the conditions at the till surface. Figure 8 shows the estimated annual sublimation rate for various till thickness, as well as the sublimation rate at each modeled time step for a maximum till depth of 70 cm. The rate of ice loss under 1 cm of till is 1.64 mm yr⁻¹, and under 70 cm, 0.066 mm yr⁻¹; these values represent the net change including sublimation and condensation (i.e., vapor flow towards the ice surface). During the model, the average D_{eff} throughout the 70 cm-thick sediment column ranged from 4.7 x 10⁻⁶ to 5.5 x 10⁻⁶ m² s⁻¹ with an average of 5.1 x 10⁻⁶ m² s⁻¹. Using the sublimation rate vs. till thickness data (Fig. 8a), we estimate the time to reach a 70 cm sublimation till cover to be ~0.4 and 4.2 Myr for 10% and 1% ice sediment content respectively.

Figure 8. Modeled sublimation rates for Mullins Glacier. (a) Average annual sublimation rates for till thicknesses of 1 cm to 70 cm. (b) Sublimation rate at each time step for ice under 70 cm of till. Time series reflects estimates of the sublimation rate at 30-minute time intervals from January 1, 2011 to December 31, 2012. A negative value represents mass loss via sublimation, and a positive value represents vapor diffusion inward towards the buried ice surface. For 24.3% of the modeled time steps, vapor was transported downward toward the ice surface (i.e., the flux was > 0).

7. Discussion

7.1 Factors influencing diffusivity and sublimation rate

Ice impurities. In order ensure that the ice used in each sublimation experiment had identical properties, DI water was used for our ice cores; however, glacier ice contains impurities including salt at grain boundaries (Cullen and Baker, 2001) and as inclusions (Ohno et al., 2005), dust (Lambert et al., 2008), and concentrated debris layers (Mackay et al., 2014) all of which will affect its chemical and physical properties. For example, a recent study by Kossacki and Leliwa-Kopystynski (2014) showed that mineral impurities affect the temperature dependence of ice sublimation rates. In future experiments, polycrystalline ice representing the measured debris and salt content of glacial ice will be used to better understand glacial sublimation rates.

Snow and boulder cover. Although we do not explicitly consider the role of snow cover in modulating the sublimation rate of buried ice through the Mullins till, snowfall events are at least partially captured in our model by our RH measurements, which approach 100% during periods of snow (see Figure 6). McKay (2009) showed that snow cover has a stabilizing effect on

sublimation of buried ice bodies, and sublimation models by Liu et al. (2015) and Kowalewski et al. (2012) show that snow cover significantly reduces sublimation rates in this region. Schorghofer (2009) showed that percolating meltwater, perhaps induced via albedo-induced melting of snow or a warmer climate, could counterbalance ice loss via sublimation and increase the duration of buried ice bodies. Additionally, the presence of large boulders on the surface and within Mullins till will decrease D_{eff} by effectively decreasing the overall till porosity and increasing tortuosity, as water vapor molecules will be forced to travel a further distance around the obstacles in order to reach the till surface. Kowalewski et al. (2012) found that the presence of a rocky pavement on the till surface decreases annual sublimation by more than 50%. Adding the effects of snow cover, meltwater, and large clasts in our experiment would most likely result in reduced sublimation rates.

Salts. The presence of nitratine in Mullins till may play a role in modulating loss of Mullins Glacier ice. Nitratine is highly deliquescent and rapidly absorbs water from the surrounding air (Steiger, 2003). High concentrations of NaNO_3 , or other deliquescent salt species, may result in water vapor becoming absorbed to salt crystals when the local RH or vapor concentration is high; a process that can be reversed (i.e., efflorescence) when the RH lowers (Hoffman et al., 2004). These processes have been observed in salt-rich soils elsewhere in the Dry Valleys (Dickson et al., 2013) and locally can alter vapor pressure gradients driving diffusion and prevent vapor from diffusing through the entirety of the soil column; efflorescence can also lead to changes in sediment porosity.

Long-exposed tills in the MDV develop a vertical salt stratigraphy whereby the concentrations of Na and SO_4 show a peak at a few cm depth (reflecting the downward percolation of snowmelt and subsequent evaporation) and a general decrease thereafter with

depth (Bao and Marchant, 2006). In addition, sublimation tills, like Mullins till, also show a secondary peak in salt concentration, typically just above the buried-glacier surface (Bao et al., 2008). Salt concentrations at all depths tend to increase over time, and for deposits with long-exposure durations (3-5 Ma) in inland regions of the Dry Valleys peak salt concentrations are typically 1-2 wt% for [SO₄] and <1 wt% [Na] (Bao and Marchant, 2006). Discrete, low-porosity salt layers would likely result in decreased vapor transport throughout the sediment column, and hence a reduction in sublimation rates.

7.2 Comparison with previous studies

Multiple studies have examined the sublimation rates of ice in the MDV, including studies of massive ground ice (e.g., Kowalewski et al., 2011; Liu et al., 2015) and ice-cemented ground (e.g., Hagedorn et al., 2007; McKay et al., 1998). Our modeled sublimation rates compare closely to previous studies that examine vapor diffusion under various sediment depths. For example, Liu et al. (2015) found annual sublimation rates between 0.17 mm and 0.09 mm for ice in Beacon Valley under 50 cm of sediment; our model predicts an annual sublimation rate under 50 cm of till to be ~0.11 mm under present-day forcing. Similarly, McKay et al. (1998) modeled the annual sublimation rate of a high elevation ice-cemented body in the Asgard Range under 25 cm of sediment to be 0.4-0.6 mm, and Hagedorn et al. (2007), found the annual sublimation rate of ice-cemented ground under 22 cm of sediment in Victoria Valley to be 0.22 mm; our model predicts sublimation rates under 25 cm of sediment to be ~ 0.31 mm yr⁻¹. We expect differences in sublimation rates between studies in the MDV based on the atmospheric forcing of each specific study site (which varies with elevation and distance from the Ross Sea), as well as the modeling approach. Additionally, the sublimation rates from a flat surface, as

modeled here, would be expected to differ from rates modeled in polygon troughs, or Many of these studies have used effective vapor diffusivity (D_{eff}) values less than those determined here (in some cases, less than half), between 1.8 and $3.1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ (Hagedorn et al., 2007; Kowalewski et al., 2011; Liu et al., 2015). McKay et al. (1998) determined a D_{eff} of $4.1\text{-}5.5 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, which is within the range of diffusivities determined in this study. It is possible that some of these disparities could be explained by site-specific differences in lithology and weathering history that would affect the grain-size distribution and other sediment properties.

8. Conclusions

This study is the first of its kind to use field-site sediment samples and a climate-controlled environmental chamber to examine sublimation rates of buried ice under Antarctic environmental conditions. We determine an effective diffusivity for Fickian vapor transport through Mullins till using experimental measurements of mass loss via ice sublimation beneath varying thicknesses of sediment. Results show D_{eff} at -10°C , 40% RH, and 850 mbar is $(5.2 \pm 0.5) \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$. We use this value, coupled with micrometeorological data prescribing local environmental conditions on Mullins Glacier, to estimate the sublimation rate of Mullins Glacier ice beneath 1 cm to 70 cm of Mullins till, the till thickness overlying the oldest portion of Mullins Glacier. The model results show that ice loss beneath 70 cm of till is 0.066 mm yr^{-1} . These experimental and modeling results support recent findings that call for multi-million year preservation of buried glacier ice in the coldest and driest regions of the McMurdo Dry Valleys (e.g., Liu et al., 2015).

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