

## Effects of Melt on Snowpack Stratigraphy and Chemistry on the Kahiltna Glacier, Central Alaska Range

DOMINIC WINSKI<sup>1</sup>, KARL KREUTZ<sup>2</sup>, ERICH OSTERBERG<sup>3</sup>, SETH CAMPBELL<sup>4</sup>, AND  
CAMERON WAKE<sup>5</sup>

### ABSTRACT

Paleoclimate and mass balance studies depend on the determination of the physical and chemical effects of melt water on the snowpack. We identify these effects by monitoring the chemical and stratigraphic changes in snowpits at 2100 meters in elevation in the wet snow zone of the Kahiltna glacier, Alaska near the onset of the melt season. The physical properties of grain size, snow hardness and density were measured within the top 1.8 meters of snowpack twice daily between May 26-June 10, 2010. We sampled deuterium ratios as well as major ion and trace metal concentrations at the same site on timescales of one day and one month to determine the changes in chemical profiles imposed by melt. We found that the mass of surface snow melted is significantly correlated ( $R^2=0.56$ ) with both snowpack density and total mass of melt layers ( $R^2=0.398$ ) within the upper 1.8 meters of the snowpack. At this site, only 53% of the melt water retained in the snow is reconstituted in melt layers, while the other 47% is distributed evenly, leading to a homogenization of the stratigraphic profile. Melting smoothes the deuterium isotopic signal with an average 8.6 per mil enrichment in the uppermost meter of snow. After one day of melt trace metals are concentrated by 300-400% in areas at just above melt layers. Studies in ice core paleoclimatology and mass balance estimates from surface lowering must account for the physical and chemical changes imposed by melt in sites experiencing similar conditions.

**Keywords:** melt layers, densification, glaciochemistry, Alaska, snow

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<sup>1</sup> University of Maine, Climate Change Institute, 5790 Bryand Hall, Orono, ME,  
[dominic.winski@gmail.com](mailto:dominic.winski@gmail.com)

<sup>2</sup> University of Maine, Climate Change Institute, 5790 Bryand Hall, Orono, ME,  
[karl.kreutz@maine.edu](mailto:karl.kreutz@maine.edu)

<sup>3</sup> Dartmouth College, Department of Earth Sciences, Fairchild Hall, Hanover, NH,  
[erich.osterberg@dartmouth.edu](mailto:erich.osterberg@dartmouth.edu)

<sup>4</sup> University of Maine, Department of Earth Sciences, 5790 Bryand Hall, Orono, ME and U.S.  
Army Core of Engineers Cold Regions Research and Engineering Laboratory, Hanover, NH,  
[seth.campbell@umit.maine.edu](mailto:seth.campbell@umit.maine.edu)

<sup>5</sup> University of New Hampshire, Department of Earth Sciences, Durham, NH,  
[Cameron.wake@unh.edu](mailto:Cameron.wake@unh.edu)

## INTRODUCTION

The understanding of melt processes on glaciers is of key importance to the fields of glaciology and climate change. The duration and intensity of melting on glaciers are controlled primarily by quantifiable atmospheric parameters (Rupper and Roe 2008; Hoffman et al. 2008; Munro and Marosz-Wantuch 2009), which are changing, leading to different melt conditions than have been experienced previously. For instance, temperature increases in Central Alaska (Stafford et al. 2000), the location of our study site, coincide with accelerated glacier melting and negative mass balances (Arendt et al. 2002; Berthier et al. 2010; Josberger et al. 2007; Luthcke et al. 2008). Whether or not this increasing volume of meltwater refreezes within the glacier or runs off into streams and oceans has important implications for estimations of glacier mass balance, the predictions of the contribution of alpine glaciers to sea-level rise, and water availability and run-off prediction.

In addition to the glaciological importance of melt, many studies have successfully used melt layer frequency or percent melt in ice cores as a proxy for summer temperature (Herron et al. 1981; Das and Alley 2005; Alley and Anandakrishnan 1995; Kelsey et al. 2010; Fisher et al. 1995). The reasoning behind this approach is that warmer summers lead to more surface melt which will eventually percolate down through the snowpack and become refrozen in discrete layers that can later be measured and placed in a chronological context. These studies are predicated upon several assumptions including: 1) melting is strictly proportional to temperature; 2) that the meltwater formed at the surface will be organized in discrete layers whose thickness is proportional to the volume of melting produced at a definable temperature event; and 3) melt layers will form and remain within that year's snowpack providing an accurate chronological record. We test these assumptions at our study site in the wet snow zone of the Kahiltna Glacier in Denali National Park, Alaska.

A variety of glaciochemical records are used for paleoclimate reconstruction in addition to ice-core based melt layer paleothermometry. This approach also relies on the fact that snow stratigraphy is laid down in chronological order, with an increase in depth equating to an increase in time. Post depositional processes such as wind transport, melting, and folding complicate the original surface record preserved in ice cores and must be thoroughly understood or avoided before interpretation. Given the increasing glacier melt throughout much of the cryosphere (IPCC 2007, Meier et al. 2007) and the fact that some ice core sites are located in zones experiencing significant melt (Grumet et al. 1998; Virkkunen et al. 2007), we attempt to identify patterns in how melt affects the physical and chemical stratigraphy preserved in the snowpack. The identification of these patterns and their application to stratigraphic records will lead to a more robust interpretation of ice cores in warm regions.

In this study, we quantify the effect of surface melting on snow stratigraphy in the wet snow zone through observations of physical snowpack properties, the formation of melt layers, and the effects on the chemical stratigraphy within the snow. Specifically, we aim to build on work done by Arendt et al. (2002), Berthier et al. (2010) and Josberger et al. (2007) by adding to the description of glacier mass balance through budgeting the fate of surface snow and ice that are melted at Kahiltna Base Camp (Kbase). In doing this, we will test the relationship found by Kelsey et al. (2010), who linked temperature to melt percent on Kahiltna Pass, at our study site, Kbase, 1000 meters lower in elevation. As temperature reconstruction from melt layers depends on a preservation of the melt layer record within the snow, monitoring the stratigraphic evolution at Kbase will test the limits of this proxy. Finally, we assess the impact of melt on the chemistry preserved in the snowpack and determine the suitability of similar regions to ice core work.

## STUDY AREA

To assess the impact of surface meltwater on a glacial snowpack, we monitored conditions at Kahiltna Base Camp in Denali National Park, Alaska. The Kahiltna Glacier is located at roughly 63.5° N and 151° W in the Central Alaska Range, immediately below the summit of Mt. McKinley (Denali). This study site was chosen primarily because of the melt conditions at the site as less is known about the preservation of ice core records in the wet snow zone compared with

the percolation and dry snow zones, although it is generally agreed that the greater the degree of melting, the more post depositional alteration will occur (Grumet et al. 1998; Taylor et al. 2001). We chose to work at KBase because the high intensity of summer melt will help us to understand how the records preserved in the snowpack change in such environments.

Kahiltna Base Camp (Kbase) lies at roughly 2100 meters in elevation on the Southeast Fork of the Kahiltna Glacier (Figure 1). As the ELA is near 1800 meters in elevation during most years (Campbell et al. 2011) Kbase is located firmly within the wet snow zone. By late May, when this study commenced, melting had already begun with temperatures being generally above freezing during the day. The site receives 1 to 2 meters of precipitation, in water equivalents, per year, much of which is lost to melt (Winski et al. 2011).

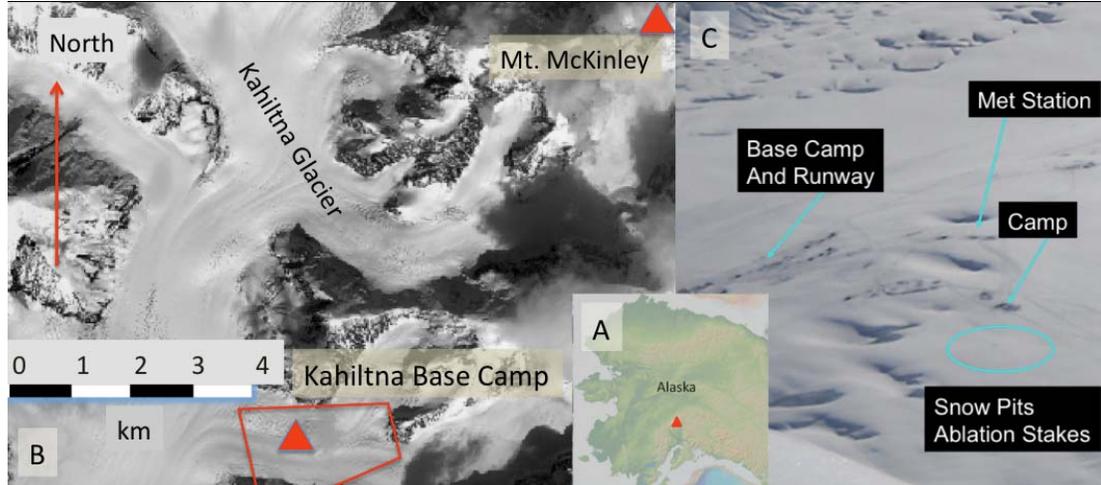


Figure 1. Location map. A) The position of our site in Central Alaska B) The Kahiltna Glacier with the location of Kahiltna Base Camp C) Photo of the study site with relevant locations.

## METHODS

### Physical Parameters

As the goal of this study is to understand the ways in which meltwater affects the physical and chemical evolution of the snowpack, we measure both the snow properties and the amount of melt over time. Two methods were used to estimate the volume of surface melt. First, five ablation stakes were placed about 30 meters from one another in a ring. This serves both to integrate any microscale spatial variation at the site as well as to keep the area pristine and untrodden by climbing parties. The data from all five stakes were averaged each day. Surface snow density measurements were taken at the time of the stake reading using a cutter of known volume (68.3 cm<sup>3</sup>) and a scale. Density measurements were taken 3 times, averaged, and used in conjunction with the surface lowering to calculate the total amount of ablation  $\dot{A}$  (kg/m<sup>2</sup>/hr or mm w.e./hr):

$$\dot{A} = (H_1 - H_2) \frac{\rho}{\Delta t} \quad (1)$$

where  $H_1$  is the snow height at time 1 in meters,  $H_2$  is the snow height at time 2,  $\rho$  is the density of the surface snow in kg/m<sup>3</sup> and  $\Delta t$  is the length of time between measurements.

### Sampling Schedule: 2010 Field Season, Kahiltna Base Camp

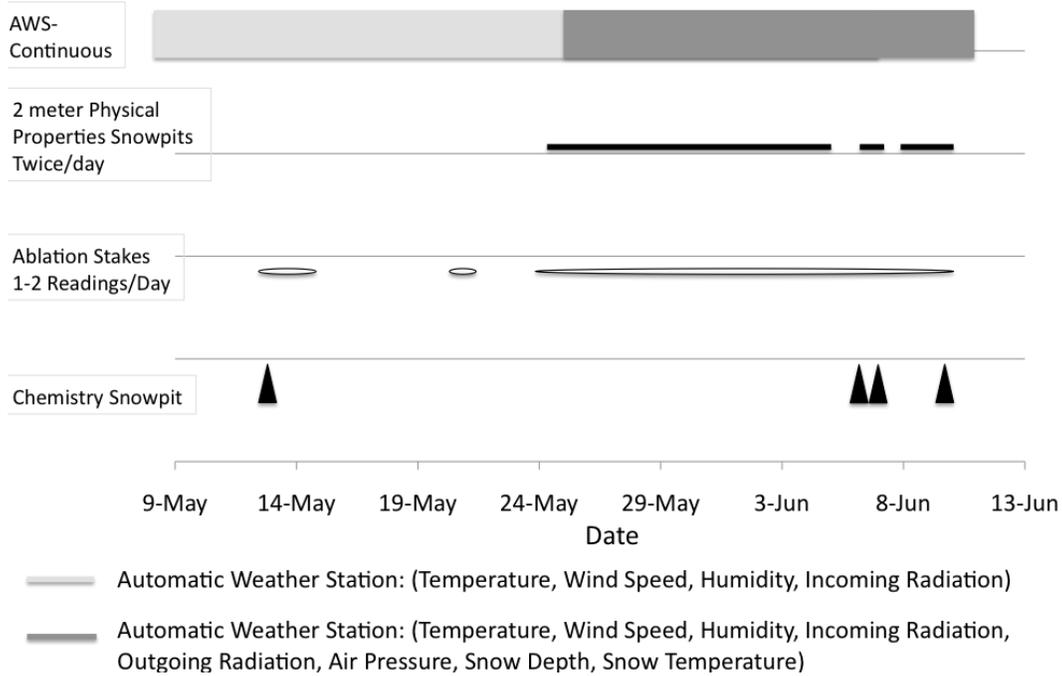


Figure 2. Schedule of measurements taken at Kbase during 2010. Partway through the season the meteorological station was improved to measure additional variables.

Not necessarily all of the snow loss is due to melting, nor does all of the meltwater infiltrate the snowpack. Some of the ablation is through sublimation and evaporation, and does not affect lower snow stratigraphy (Gusain et al. 2009; Stichler et al. 2001; Taylor et al. 2001). Water loss to the atmosphere affects isotope ratios differently than meltwater percolation (Taylor et al. 2001; Stichler et al. 2001). Also, water lost to the atmosphere does not form melt layers, affect lower snow stratigraphy or move the chemistry as meltwater does. Finally, sublimation and evaporation are an unrecoverable loss to the mass balance of the glacier, whereas water melted at the surface may refreeze within the snowpack (Stichler et al. 2001; Conway and Benedict 1994; Pfeffer and Humphrey 1996).

We calculated the surface energy balance at Kbase to distinguish sublimation and evaporation from melting. To do this, we used an automatic weather station (AWS) to collect the following data each hour: air temperature ( $^{\circ}\text{C}$ ), wind speed (m/s), incoming solar radiation ( $\text{W}/\text{m}^2$ ), relative humidity (%) and snowpack temperature ( $^{\circ}\text{C}$ ). From the meteorological data, we constructed a surface energy balance model (Winski et al. 2011) to calculate the radiative and turbulent heat fluxes. These energy fluxes were summed to give an hourly energy balance for the snow surface which, using the latent heat of fusion of water, can be converted to an ablation rate (Brock and Arnold 2000; Anslow et al. 2008; Klok et al. 2005; Hock 2005; Schneider et al. 2007).

The AWS data allow us to separately calculate the energy fluxes due to sublimation and evaporation showing that only 2% of the snow is lost to these processes while 98% enters the lower snowpack as melt. The values for ablation produced by the energy balance model are then scaled by the ablation stake data so that the total amount of melting over the study period computed by the model equals that measured at the stakes.

We assume that all melt is produced at, or very near, the surface. Studies have shown that shortwave radiation can penetrate a significant depth and cause subsurface warming and melting, especially in blue ice areas (Brandt and Warren 1993). However, the presence of meltwater leads

to the attenuation of incoming radiation in the snowpack (Kapil et al. 2010). Because of the high absorptivity of the snow on the Kahiltna glacier and the previous success in using surface energy balance to calculate melting (Anslow et al. 2008; Konya and Matsumoto 2010; Schneider et al. 2007) we feel confident that the vast majority of the meltwater at this site is generated at the surface.

To measure how the physical properties of the upper snowpack change with the addition of meltwater, a 1.8-meter snowpit was hand dug twice a day over the duration of the study period (Figure 2). Pits were dug in a seldom traveled area that had been protected by flags and wands for 2 weeks preceding sampling. All sampling faces were oriented so as not to receive direct sunlight and were constructed upslope and one to two meters away from a previous pit or disturbed area.

We measured snow hardness, grain size, and density each time we dug a snow pit. Hardness was measured semi-quantitatively using the method described in International Classification for Seasonal Snow and the Ground (Colbeck et al. 1990). The field measurements and the corresponding hardness values are given in Table 1.

**Table 1 The snow strength was assigned to each layer in the stratigraphic column based on the ability of the objects, listed at left, to easily penetrate the snow (Colbeck et al. 1990).**

Largest Object Able to Easily Penetrate Snow	Snow Strength (Pa)
Fist	500
4 Fingers	5000
3 Fingers	10000
2 Fingers	20000
1 Finger	50000
Pencil	500000
Knife Blade	5000000

Melt layers, being any icy lens or large rounded grains held by a matrix of ice, were also recorded at this time. The grain size and density were sampled, not at a constant resolution, but based on the layers identified by hardness. Grain size was calculated by scraping the pit sides and recording the grain diameters based upon measurements by a ruler. Density was measured for each stratigraphic layer using a cutter of known volume and a scale. Measurements were repeated three times and averaged.

### Chemistry

We assessed the impact of melt on snow chemistry on timescales of one day and one month. Upon arriving at Kbase on May 13, 2010 a 3.15 meter snowpit was dug in an undisturbed area close to the AWS followed by the extraction of a 6.6 meter core from the pit bottom. The pit was scraped with clean plastic scrapers, the stratigraphy was recorded, and a series of snow samples were taken from the pit wall. Each sample was 15 cm thick and the samples provided a continuous profile of the pit wall. Samples were taken for  $\delta D$ , major ions ( $Na^+$ ,  $K^+$ ,  $Cl^-$ ,  $Mg^{2+}$ ,  $Ca^{2+}$ ,  $MS$ ,  $NO_3^-$ ,

SO<sub>4</sub><sup>2-</sup>), and 26 different trace metals (Sr, Cd, Cs, Ba, La, Ce, Pr, Pb, Bi, U, As, Li, Al, S, Ca, Ti, V, Cr, Mn, Fe, Co, Na, Mg, Cu, Zn, K) and placed in Nalgene HDPE 175 ml plastic bottles. Before sampling, bottles for  $\delta$ D and ions were repeatedly soaked and washed in de-ionized (DI) water while those for trace metals were washed and soaked in 10% trace metal grade nitric acid and then cleaned with DI water according to the Climate Change Institute chemical analysis procedure. Sampling was conducted while wearing Tyvek clean suits and polyethylene gloves similar to protocols in Fortner et al. (2009) and Goto-Azuma et al. (2006).

On June 10, another 2.85 meter snowpit was dug within 3-4 meters of the initial pit so as to have the same stratigraphy while not experiencing any thermal alteration from the opening of the initial pit (Figure 2). The second pit was sampled in the same manner as the initial pit and for the same chemical constituents in order to see whether melt during the previous month had altered the chemical profile.

A similar experiment was done on the timescale of a day. Early in the morning of June 7, a day forecasted to be warm and sunny,  $\delta$ D, major ions, and trace metal samples were taken from that morning's physical properties pit (Figure 2). The same sampling and clean procedures were followed as in the case of the two monthly pits. Sampling took place at 10 cm resolution to a depth of 1.3 meters. The procedure was repeated that evening after the bulk of the day's melting had taken place.

Samples were analyzed at the University of Maine Climate Change Institute.  $\delta$ D was measured in melt water from discrete layers in each of the 4 chemistry pits (111 samples) by gas isotope ratio source mass spectrometry. We used Cr reduction techniques (Morrison et al. 2001) in a Eurovector elemental analyzer and a Microprime Isoprime mass spectrometer to obtain the hydrogen isotope ratios (precision +/- 0.5%). Each sample was run three times with the average of three runs being the reported result. Results are given on the SMOW scale ( $\delta$ D relative to seawater).

We used a Dionex ion chromatograph with a Gilson Liquid Handler autosampler to measure concentrations of the eight major ions. Procedures are similar to those used by Osterberg et al. (2006).

Trace metals were analyzed using the University of Maine Thermo Electron Element2 inductively coupled plasma mass spectrometer. Samples were spiked with 1% nitric acid a month prior to analysis to insure the dissolution of all metals. All bottles and material in contact with the samples were cleaned with trace metal grade nitric acid and samples were run with Optima HNO<sub>3</sub> (Osterberg et al., 2006).

## RESULTS

### Physical

Between the evening of May 26 and the evening of June 10, ablation stakes show a total of 289 mm w.e. of ablation. The energy balance model calculated that 97.7% of this ablation was in the form of melt, so 283 mm of meltwater entered the snowpack during this time. Melt is highly variable with rates ranging from 2.5 mm w.e./hr on sunny days to 0 mm w.e./hr on cold or stormy days (Winski et al. 2011). Because air temperature was consistently above 0°C, at least the top 1 meter of the snowpack remained isothermal throughout the study period with the possible exception of the immediate surface after cold nights.

By monitoring the movement of individual snow layers we see that the stratigraphy moved toward the surface over time as each layer became shallower and, eventually, was exhumed as the surface melted away (Figure 3-5). Over the study period, the surface migrated downward a total of 37.5 cm, although the rate was extremely variable. Surface lowering, as measured by ablation stakes, is correlated with cumulative energy balance ( $R^2=0.61$ ), as is the vertical movement of internal layers.

A few features stand out clearly from the snow hardness data (Figure 3). Overall hardness increases with time, reflecting the increasing number of melt layers and the shrinkage and destruction of soft powdery layers. By the end of the experiment, none of the original powdery layers ( $H < 500$  Pa) are intact aside from the fresh surface snow.

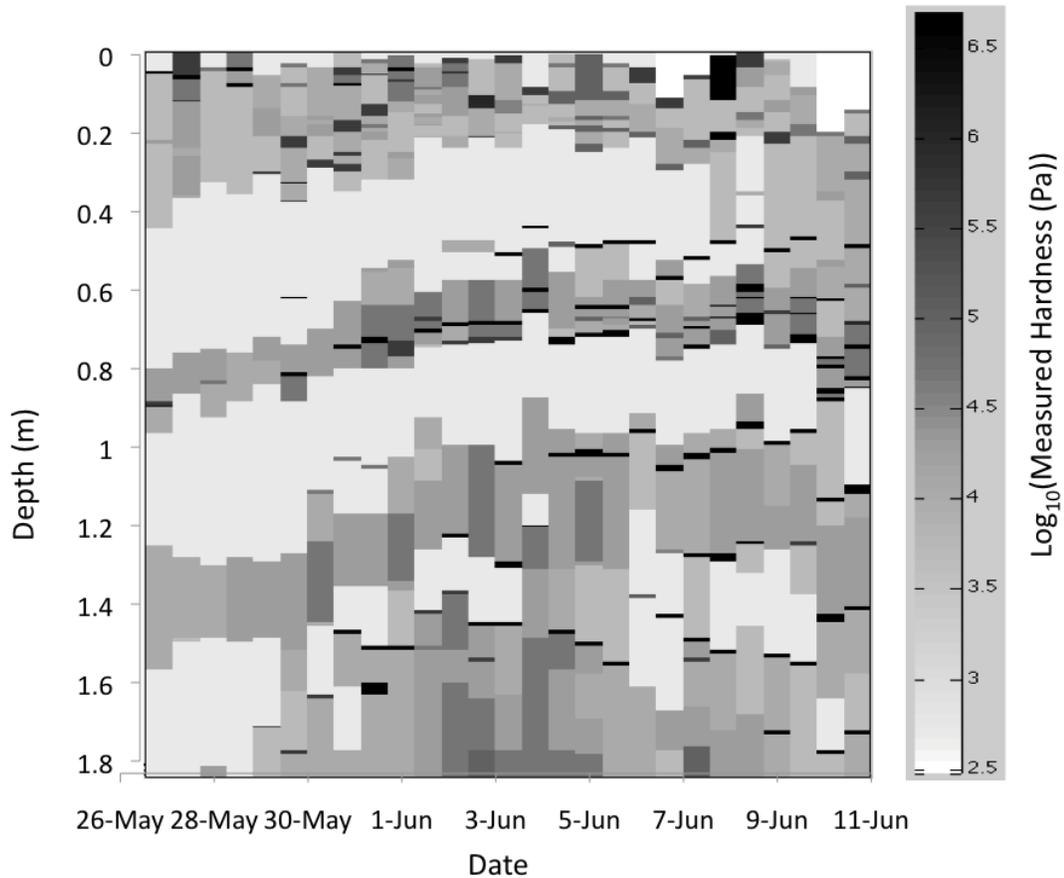


Figure 3. Evolution of snow hardness in Kbase pits over time. The grayscale represents the  $\log_{10}$  of the hardness, in Pascals, of the snow as measured by hand.

A plot of grain size over depth and time is shown in Figure 4. The banding seen in Figure 3 is reflected here as well in alternating layers of large grained soft snow and small grained hard snow. Figure 4 reveals that melt layers most frequently occur at grain size transitions. As the melt season intensifies, crystals in large grained layers tend to shrink, while those in fine-grained layers tend to grow. This results in a much more homogeneous profile at the end of the season than at the beginning.

Surface grain size increases after each storm event, but with an important exception. During the first half of the study (May 26-June 2), when there were no storms, surface grain size was neither stable, nor increasing, but fluctuated seemingly randomly by a factor of 3 (Figure 5). We suggest that under conditions when there is significant melting and no new snowfall, the principal control on the grain size will be the exhumation of previous stratigraphy. As the surface melts away and new layers are exhumed, the grain size of the surface will assume that of lower layers. This has implications for energy balance modeling or any study assuming that snow surface properties are atmospherically controlled.

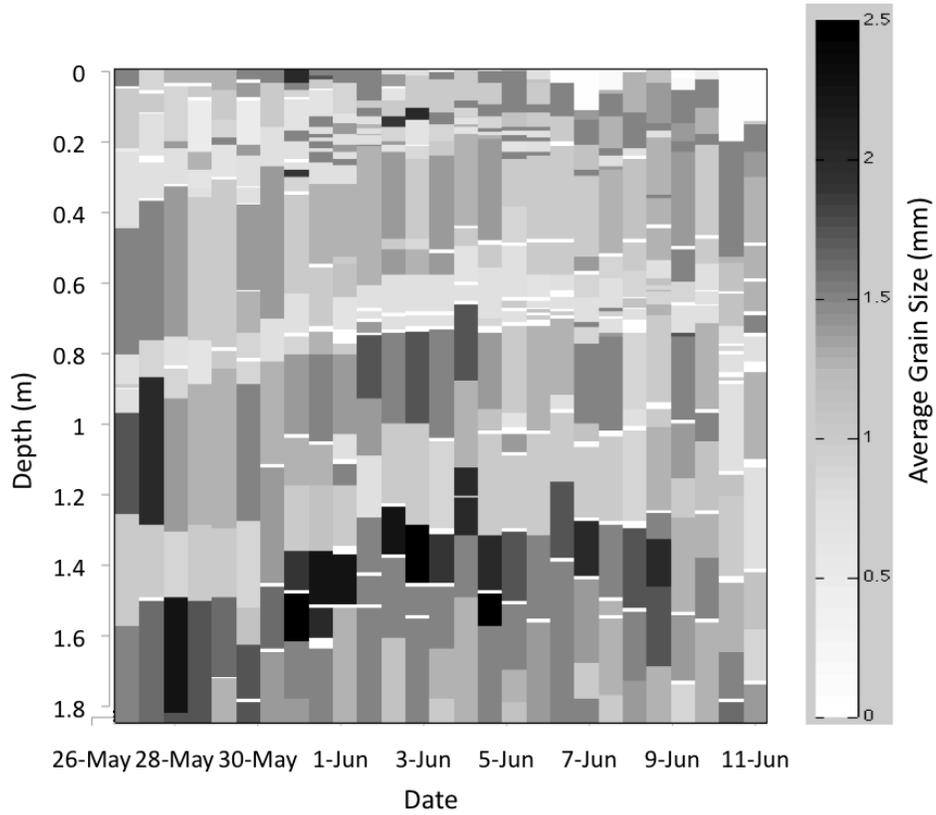


Figure 4. Evolution of snow grain size of Kbase pits over time.

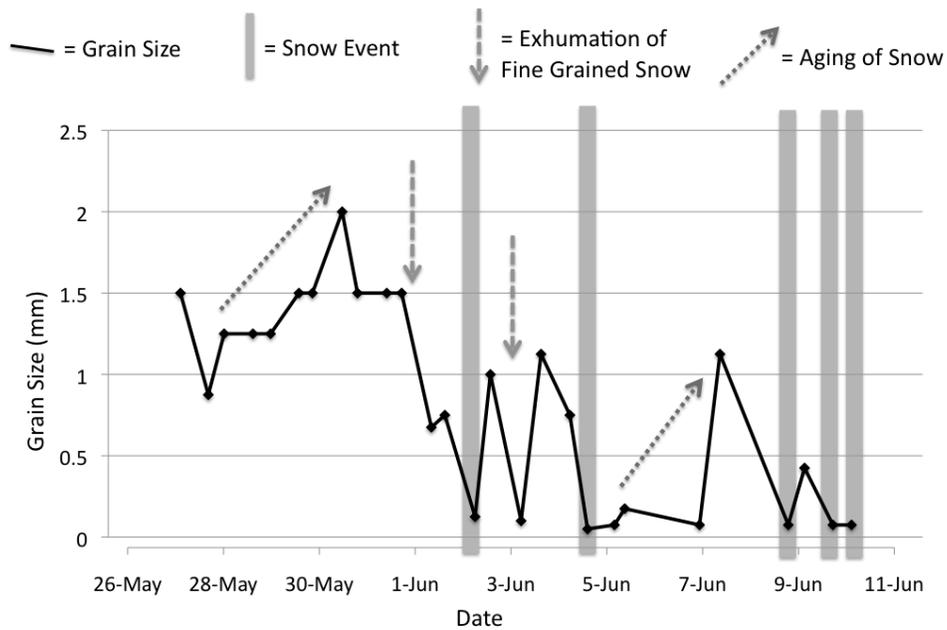


Figure 5. Surface grain size vs. time at Kbase. Surface grain size increases over time, as predicted by the literature, except in two cases: 1) a snow event that deposits very fine grained snow at the surface 2) the exhumation of a buried layer of fine grained snow.

Density of the snowpack over time is shown in Figure 6. Less dense layers tend to have large grains and be softer although, as opposed to grain size and hardness, there is a very distinct trend of densification over time. At Kbase, the mass of a 1 m<sup>2</sup> area, 1.8 meters deep, rose from 955 kg to 1142 kg between May 26 and June 10, 2010. These values were calculated by multiplying the density of each layer by its thickness and summing over the entire column. In Antarctic snow, Arthern et al. (2010) and Bell et al. (2008) observed temperature dependent densification. As we believe the mechanism for densification at Kbase is meltwater introduction, suggest a linear relationship for densification based on melt:

$$D = 0.56M \quad (2)$$

where M is melt introduced to the snowpack as calculated by the SEBM in kg/m<sup>2</sup> and D is densification in the top 1.8 meters of snow in kg/m<sup>2</sup>. The site constant for this region is a unitless 0.56, which will likely change based on region. This relation is strong for the 16 day study period (R<sup>2</sup>=0.56) but fails on diurnal timescales, showing that the densification process operates more slowly than the melt process.

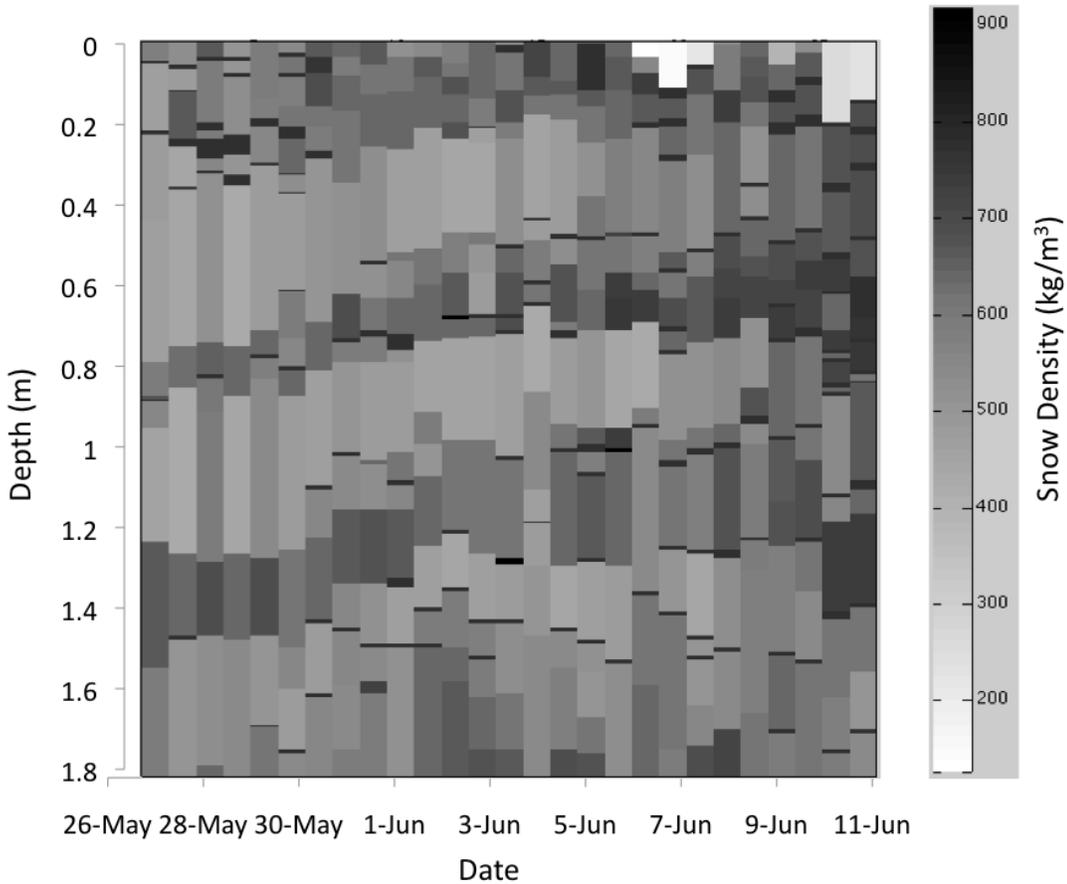


Figure 6. Evolution of snow density of Kbase pits over time.

The density data allow us to create a budget for the snow lost at the surface (Figure 7). We know from the ablation stake and energy balance data that about 289 kg/m<sup>2</sup> of snow was lost over 16 days. The upper 1.8 meters of snowpack gained about 183 kg of mass over the same period. One hypothesis for this increase is that the snow being removed from the surface is of lower density than the firm entering the profile from below (Figure 6). However, barring fresh snow events, both the average surface density and average bottom density are 600 kg/m<sup>3</sup> (+/- 5 kg/m<sup>3</sup>), which means that the mass added to the column comes from the addition of surface meltwater as

opposed to differential densities between the top and bottom of the snowpit. We also conclude that densification from settling and compaction was minimal because if the snow were being compacted significantly, the internal layers would thin and move toward the surface faster than the rate of surface lowering. Instead we observed that during the study period, the surface lowered 37.5 cm while most internal layers only moved between 10 and 33 cm while maintaining their thickness. We conclude that the 183 kg of mass added to the snowpack over the study period is principally due to meltwater addition from the surface. Of the remaining snow lost, 6 kg evaporates or sublimates leaving 100 kg (or 35%) of melt water to sink to a depth lower than 1.8 meters or exit through crevasses (Figure 7). Both scenarios are likely, as some of the snowpits ended in crevasses at or before this depth.

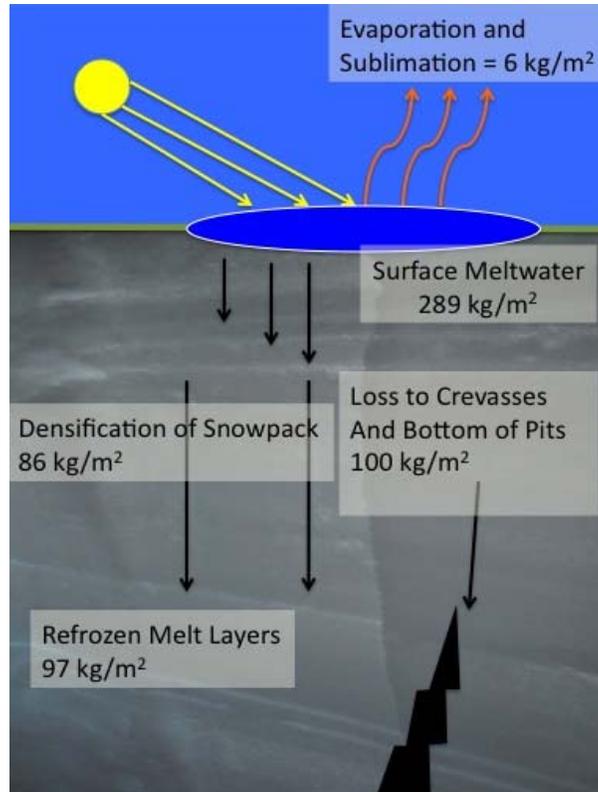


Figure 7. Fate of surface meltwater between May 26 and June 10, 2010 at Kbase. Of the 289 kg/m<sup>2</sup> of ablation at the surface, 6 kg evaporated or sublimated, 97 kg refroze in melt layers, 86 kg uniformly densified the snowpack and the remaining 100 kg infiltrated to a depth greater than 1.8 meters.

Of the 183 kg of the meltwater that remained in the upper snow column, about 97 kg or 53% went into discrete melt layers. The total mass of melt layers in the snow column increased consistently from 16 kg on May 26 to 113 kg on June 10, which is significantly correlated ( $R^2=0.40$ ) with the total energy balance and cumulative surface melt. The remaining 86 kg of water went into increasing the density of softer layers in the snowpack.

There are also distinct patterns to the location where melt layers formed within the snow pack. On May 26 only 3 thin melt layers were present in the snow column; by June 10 there were 12 layers of varying thickness. Most melt layers formed at transitions in snow hardness, density, grain size, or a combination of these (Figures 3-6). This is may be due to differences in permeability among snow types leading to water buildup, and later refreezing, at transitions. However, this theory suggests that melt layers should form preferentially at transitions from small grains above to large grains below (Kapil et al. 2010; Pfeffer and Humphrey 1996; Bell et al. 2008), but in fact we do not observe this. Parry et al.'s (2007) observations are similar to ours in that they did not see a preference for melt layer formation at grain size transitions. Some of the thickest and most solid melt layers do seem to have been formed around the layer of depth hoar with sharp density

decrease and grain size increase as would be expected by Pfeffer and Humphrey (1996) and Bell et al. (2008).

A plot of the grain size difference between the top and bottom of melt layers, versus the density difference on either side of the same melt layers is informative (Figure 8). First, it shows that melt layers form along transitions in both directions of grain size and density. Secondly, it shows the negative correlation ( $r^2=0.20$ ,  $p<0.001$ ) between grain size and density: when there is a drop in grain size, there is usually an increase in density. Finally, the plot delineates two different types of melt layers. The main sequence of melt layers is found at moderate grain size and density transitions within the snowpack. The other class of melt layers form after a fresh snow event, since the base of the new snow is characterized by a sharp increase in both grain size and density. Easily melted, the new snow produces meltwater, which is then refrozen at this transition.

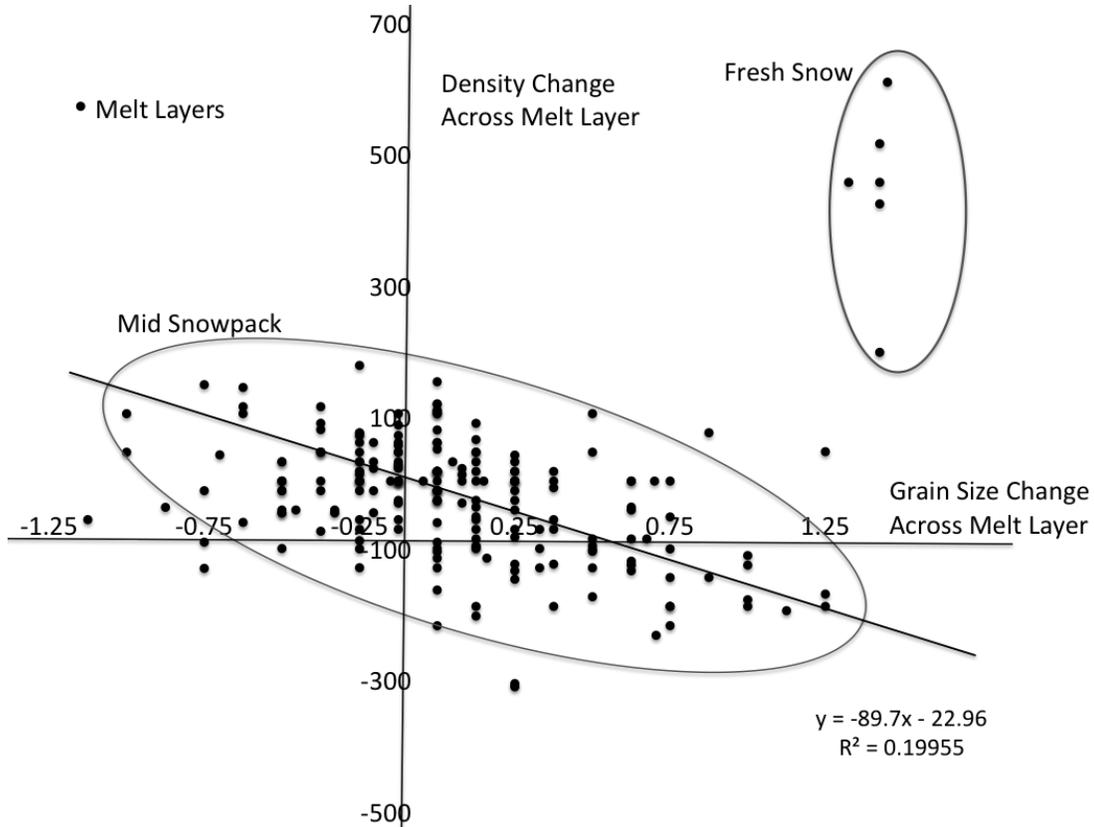


Figure 8. Melt layers recorded in the study, plotted by the difference in grain size from top to bottom of the melt layer, versus the difference in density over the same interval. Positive values indicate increasing density and grain size downward. The two principle types of melt layers are shown on the plot as the regression line for the mid-snowpack melt layers.

Similar to Moran and Marshall (2009) and Campbell et al. (2005) we observed severe homogenization of the snow profile during progressive stages of melting. During the final snow event, nearly all the identifiable snow layers either disappeared or became less distinct despite very little melt occurring at this time. It may be that the melt flow at this point changed from the restricted melt layer controlled flow to more uniform matrix flow as observed by Bell et al. (2008) during the late summer. We infer that snow events and the subsequent melting of the fresh snow tend to have a homogenizing effect on the snow profile.

### Chemical

The major feature in the snowpits from May 13 and June 10, 2010 was a 1 cm thick layer of volcanic ash, which we presume to be from the 2009 Mt. Redoubt eruption (Figure 9). The depth to the layer on May 6, 2009 was only 46-50 cm (Osterberg, personal communication). In 2010,

this layer was visible at 2.25-2.3 meters depth in both the initial and final pits. This implies 1.75-1.8 meters of accumulation, a realistic value for the site. Additionally, spikes in most trace metals are orders of magnitude higher than the background for most species, leading us to conclude that we are, indeed, seeing the 2009 Redoubt ash horizon. Sulfur and strontium data from both field seasons are plotted as examples of the ash horizon's prominence in the chemical stratigraphy, though spikes in nearly all trace metal samples were seen at the position of the volcanic peak. Trace metal concentrations vary by species but spikes at the layer itself are an average of 47 times the background level of trace metals.

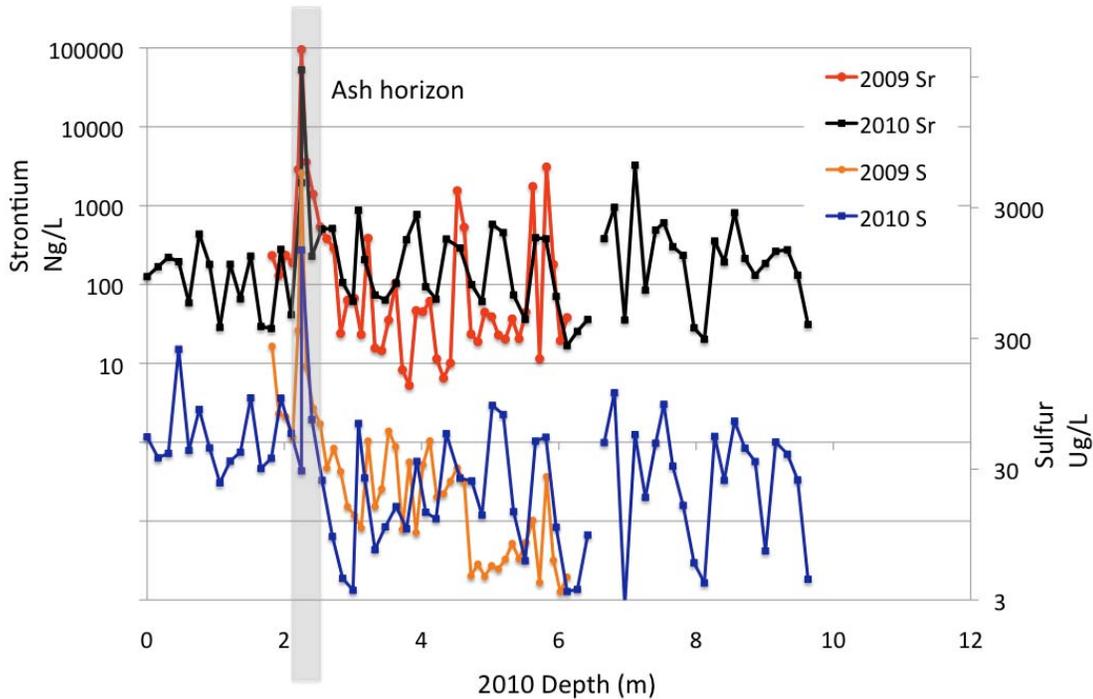


Figure 9. Sulfur and strontium data from 2009 and May 2010 snowpits at Kbase. Data are plotted on the May 2010 depth scale (1.77 meters were added to the 2009 depths). These data show the prominence of the signal from Mt. Redoubt in the snowpits from both years.

Isotopes: We sampled  $\delta D$  in the morning and evening of June 7, at 10 cm intervals in the top 1.3 meters of a snow pit (Figure 10). A total of 19 mm of melt at the surface occurred between the sampling times.  $\delta D$  of the snow was 1.7 ‰ higher in the evening than in the morning but with variations between 9.9 and -8.4 ‰ for individual layers.

To quantitatively measure the alteration of chemical species we used the formula:

$$\xi = \frac{\overline{Final} - \overline{Initial}}{\overline{Initial}} \quad (3)$$

where the initial and final inputs are the average concentrations of each species within the top 2.1 meters (above the ash layer, taken as background).  $\xi$  is the fractional change from background of that species. Over the course of one day  $\delta D$  had a  $\xi$  value of 0.007, which is the lowest in the study. The persistence of the  $\delta D$  profile over the course of one day is not only important in terms of describing the behavior of isotopes, but it also shows that we can make comparisons between the two pits in a meaningful way. If the curves in figure 10 did not match as closely, we could not be sure whether spatial differences in the stratigraphy would confound our attempts to compare layers in the afternoon pits with those in the morning. The match between the curves shows that we are, indeed, measuring the same snow layers in each case.

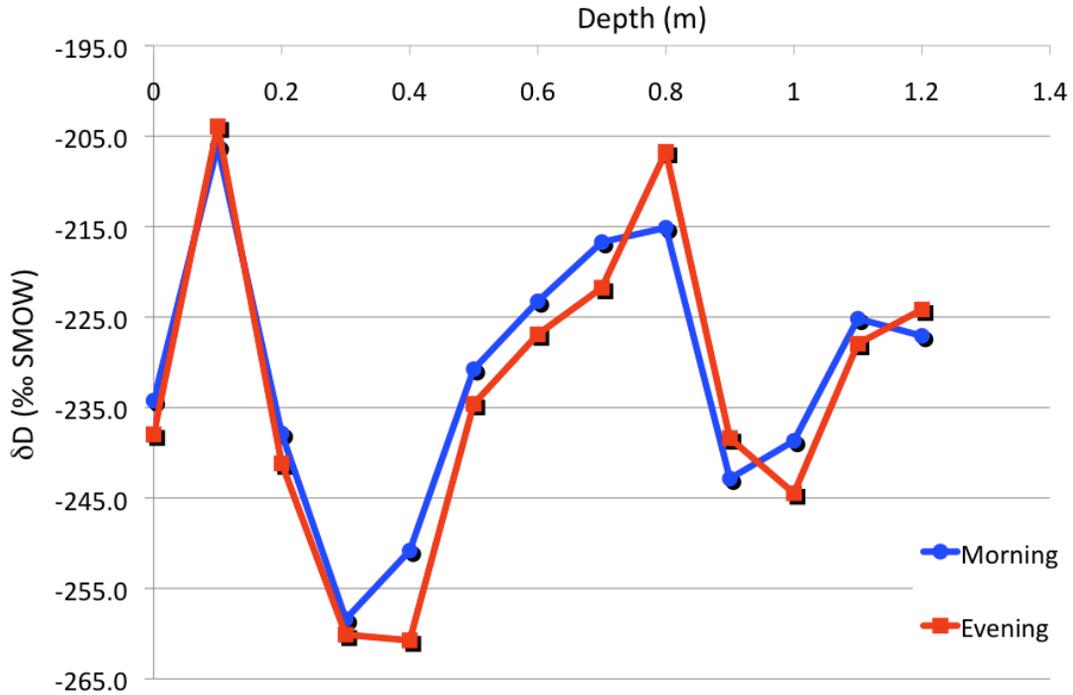


Figure 10. Plots of  $\delta D$  in adjacent snowpits in the morning and evening of June 7, 2010 at Kbase.

We compared  $\delta D$  changes over timescales of one month as well. In order to compare the stratigraphy from the pit dug on May 13, 2010 to that from June 10, 2010, we adjusted the depth scale of the June 10 pit to compensate for the 37 cm of surface melt occurring between the digging of the pits. To match the sampling depths in the two pits, we used our knowledge of surface lowering (37 cm), the location of the ash layer, and major peaks in the  $\delta D$  profiles to linearly scale the depth intervals in the June 10 pit such that they correspond to depths in the May 13 pit.

Most peaks and features in  $\delta D$  are preserved even after one month (Figure 11). As noted by Moran and Marshall (2009), Grumet et al. (1998) and Pohjola et al. (2002), the isotopic peaks are shorter and less pronounced after the onset of melting although average  $\delta D$  across the profile remains stable ( $\delta = -0.009$ ). After a month, the surface has higher values of  $\delta D$ , while lower depths have lower  $\delta D$ . Our hypothesis is that during melting and sublimation  $^1H$  is preferentially converted to liquid and gaseous water while more  $^2H$  is left behind in the upper snowpack. The meltwater, now depleted in  $^2H$ , percolates downward, depositing more  $^1H$  and less  $^2H$  lower in the profile. Heavier isotopic values in older, heavily melted or sublimated snow, have been observed in other studies such as Stichler et al. (2001), Taylor et al. (2001) and Moran and Marshall (2009), and these results show that this conclusion also applies to sites similar to the wet snow zone of the Kahiltna Glacier. However, as temperatures were generally above freezing during the period between digging the two pits, isotope diffusion is probably responsible for some of the change.

Below the ash layer is a region in which the isotopic profile has nearly uniform values of  $\delta D$  at  $-184 \pm 2 \%$ . Our hypothesis is that the deposition of the Redoubt ash layer in 2009 lowered the snow albedo enough to greatly increase the melt intensity. The quantity of melt introduced during the 2009 summer, owing to reduced albedo was sufficient to homogenize the original isotopic profile. This kind of isotopic homogenizing leads to difficulties and large errors in dating with isotopes (Moran and Marshall 2009).

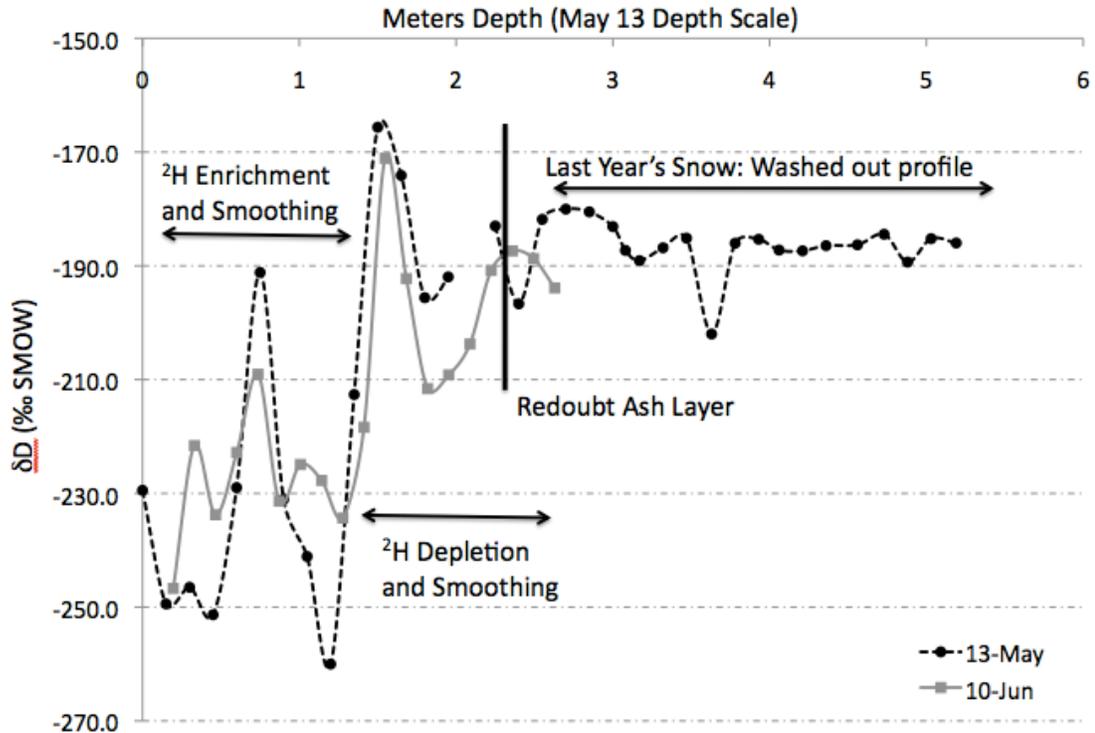


Figure 11. Changes in deuterium profile at Kbase after one month. Peaks were subdued with slight increases in  $\delta D$  in the uppermost meter and  $\delta D$  decreases below that. Below the ash layer the  $\delta D$  profile was almost completely homogenized, likely owing to the increase in melt from the lowered albedo at this time.

#### Trace Metals

The profile of trace metals is dominated by the volcanic layer in both the initial (May 13) and final (June 10) pits. Selected trace metal data from the monthly pits is shown in Figure 12. The final pit shows consistently larger spikes in concentration at the ash layer than the initial pit. This could be due to concentration at the interface between the soft snow above the ash layer and the very hard dense snow below. Also likely, is that the ash layer in the final pit fell within the center of the sampling interval, while the ash layer in the initial pit fell on the very edge, leading to lower concentrations collected in the initial sample.

Like the ash layer itself, the background level of trace metal content in the pit was almost universally enriched in the near surface pits over one month with only S, Na and Ca showing drops in average concentration. The magnitude of  $\xi$  in the trace metals ranged from 0.04 in lead to 2.05 in aluminum, though the high  $\xi$  value for aluminum was mostly due to a single spike in the second pit. Other metals whose signals are resistant to melting are cadmium,  $\xi=0.1$ , promethium,  $\xi=0.09$ , and lithium,  $\xi=0.12$ . A month in this type of climate is sufficient time for introduced meltwater to change the chemical signal significantly, but not beyond recognition, for most species.

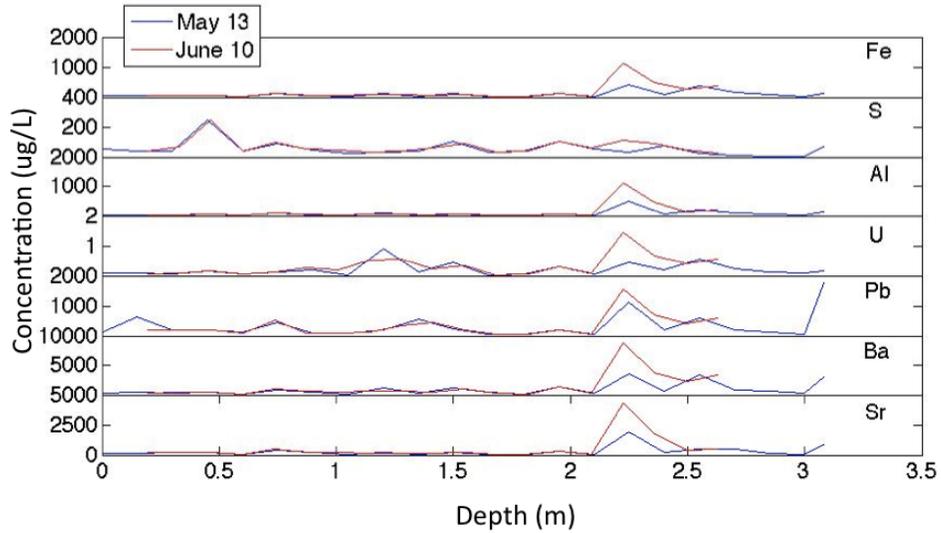


Figure 12. Changes in concentration of selected trace metals at Kbase over the course of a month.

Trace metal alteration over one day was not as strong as in the monthly pits, but was nevertheless significant. Unlike the monthly pits,  $\delta$  varied from V, at -0.54, to Ca, at 0.01, to As, at 0.59 with Pb, Bi and Li also being relatively high at 0.43, 0.43 and -0.49, respectively. Despite the visible patterns of trace metal concentration (Figures 13-14), there is no positive trend in  $\delta$  as seen in the monthly pits.

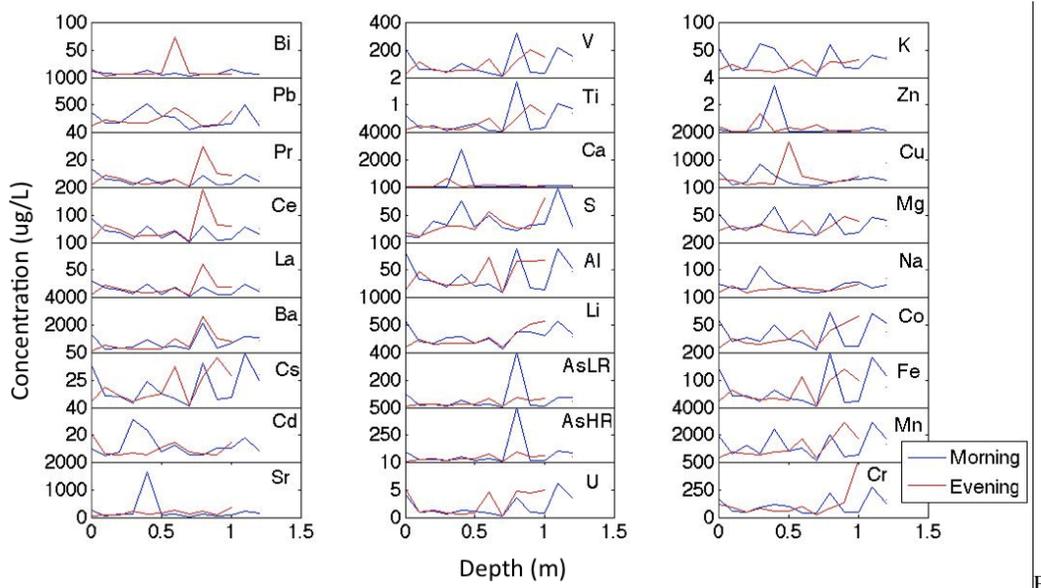


Figure 13: Trace metal concentrations at the beginning and end of June 7. Significant redistribution of trace metals is seen over the course of the day. Though many metals have similar behaviors, others behave uniquely.

Taking the ratio of final to initial concentration for each layer over the course of the day shows where trace metals are enriched with melt. Most trace metals exhibit similar trends having relatively stable levels in the upper 0.5 meters of the snowpack, with depletions in the heavily melted top layer and the soft snow layer around 30 to 40 cm depth. The regions slightly above the two most well defined melt layers (about 0.7 and 1 meter depth) show a very strong enrichment of up to 400%. This shows that meltwater transports the trace metals as far as an impermeable melt

layer above which the trace metals are deposited and concentrated. By averaging the enrichments of all the trace metals together we can see this trend even more clearly (Figure 14).

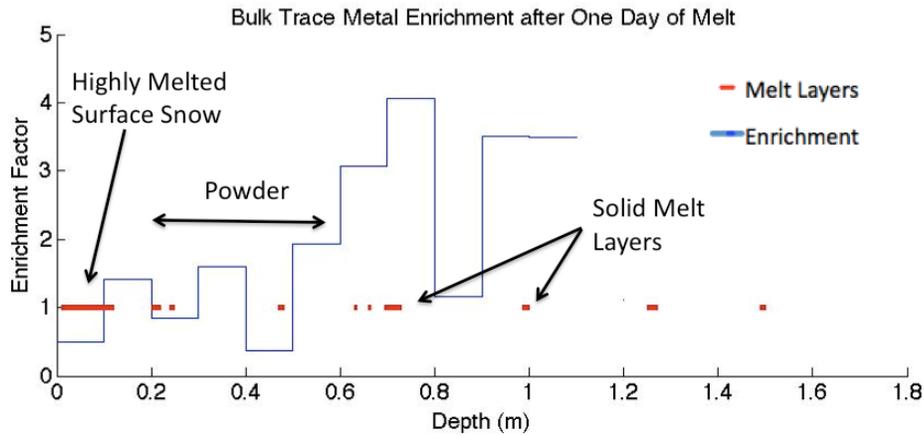


Figure 14. Relative changes in the bulk trace metal content with depth after one day. The maximum enrichments occurred at or immediately above melt layers, while surface and powdery snow was depleted, demonstrating the downward transport and easily mobilized nature of the trace metals at this site.

#### Major Ions

Changes in the ionic chemical profiles were observed between both the monthly (Figure 15) and daily (Figure 16) pits. For the monthly pits, the ions  $\text{NO}_3^-$  and  $\text{SO}_4^{2-}$  seem to be the most sensitive with an average change in concentration of 37 and 64  $\mu\text{g/L}$  respectively. This agrees with the results of Bales et al. (1989) who also found  $\text{NO}_3^-$  and  $\text{SO}_4^{2-}$  to elute most strongly, with  $\text{SO}_4^{2-}$  being affected first.  $\text{Ca}^{2+}$  was next most sensitive with a change of 18  $\mu\text{g/L}$  in average concentration. The rest of the ions all changed an average of less than 8  $\mu\text{g/L}$ .  $\xi$  is highest in the acids and divalent cations with  $\text{SO}_4^{2-}$  being the highest at -0.46. The monovalent ions are lowest with  $\text{Cl}^-$  with a  $\xi$  of -0.081. Except for  $\text{K}^+$ ,  $\xi$  has a negative value for each of the ions, indicating loss from the snowpack with melt.

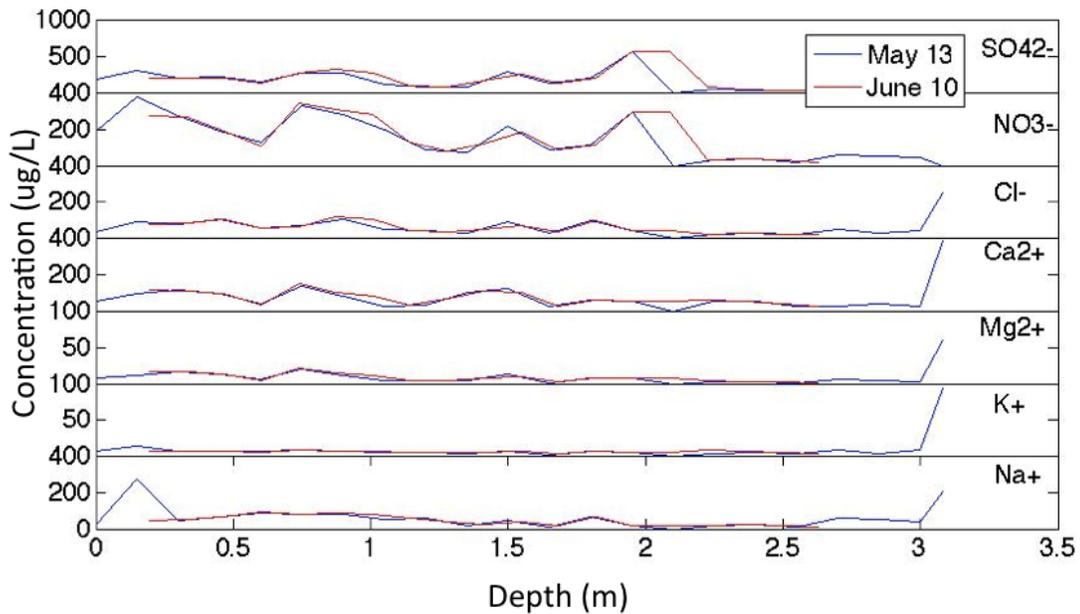


Figure 15. Changes in concentration in major ions at Kbase over the course of a month. Most major features are preserved after one month of melt.  $\text{SO}_4^{2-}$  and  $\text{NO}_3^-$  show the most susceptibility to downward transport.

The ionic signal from the snowpack was mostly preserved over one day, though not nearly to the extent of deuterium ( $\xi = 0.007$ ) (Figure 16).  $\text{Na}^+$ ,  $\text{K}^+$ , and  $\text{Cl}^-$  remain reasonably stable with spikes in concentration moving slightly deeper in the snowpit. Magnesium, sulfate and nitrate show a different story with each displaying a large spike at 0.1 meters depth and a smaller one at about 0.5 meters depth. The 0.1 m spike shrinks and the spike at 0.5 meters grows, suggesting that some of the ions were translocated downward from one peak to the other. As expected,  $\text{Ca}^{2+}$ ,  $\text{Mg}^{2+}$ ,  $\text{NO}_3^-$ , and  $\text{SO}_4^{2-}$  had the highest magnitude alteration index with  $\xi$  being 0.27, 0.5, 0.25 and 0.37 respectively, although the positive values contrast with results from the monthly pits. The monovalent ions had close to neutral  $\xi$  values;  $\text{Na}^+ = -0.07$ ,  $\text{K}^+ = 0.09$ ,  $\text{Cl}^- = -0.07$ . Concentrations of MS were too low to draw any conclusions.

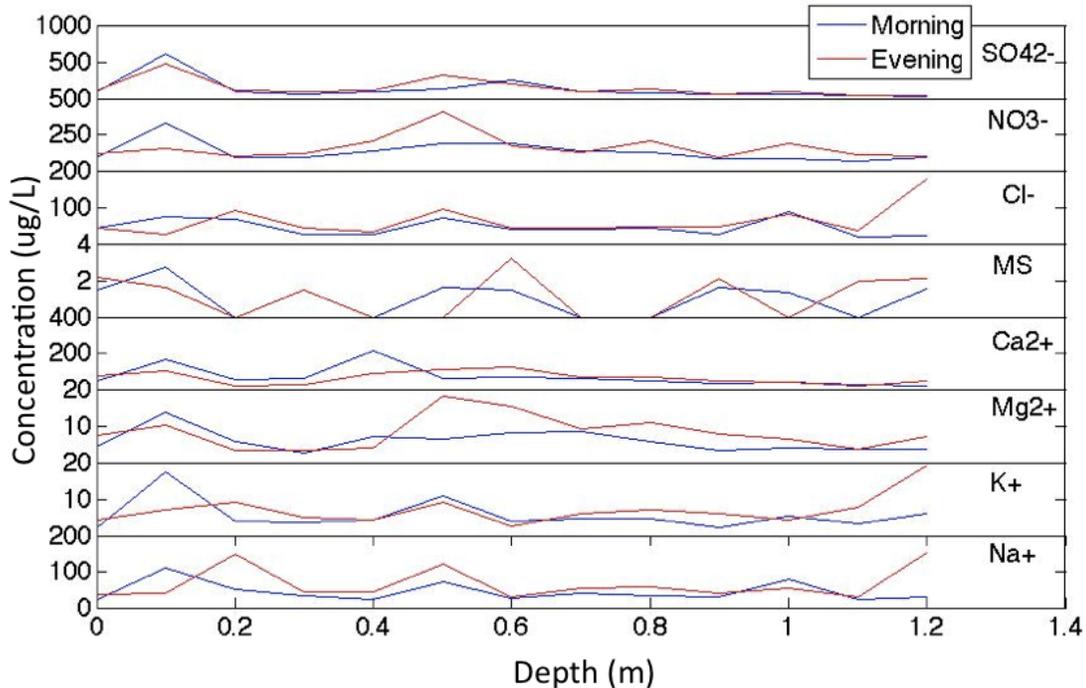


Figure 16. Ion concentrations at Kbase during morning and evening of June 7. For most ions, concentrations are eluted from near surface peaks and deposited lower. This can happen through the translocation downward of a single peak ( $\text{Cl}^-$ ,  $\text{K}^+$ ,  $\text{Na}^+$ ) or transfer of ions from a near surface peak to a pre-existing lower peak ( $\text{SO}_4^{2-}$ ,  $\text{NO}_3^-$ ,  $\text{Mg}^{2+}$ ).

## DISCUSSION

The quantity of melt occurring at Kbase as well as the energy balances recorded are well within normal ranges reported in the literature. Most of the energy available to melt the snow comes from solar radiation with the next largest source being sensible heat. This is in agreement with the work of others (Klok et al. 2005; Munro and Marosz-Wantuch 2009; Anslow et al. 2008; Gusain et al. 2009; Sicart et al. 2008; Braun and Hock 2004). The corroboration and tuning of an energy balance model with data based on measurements of stakes is also not new (Brock and Arnold 2000; Gusain et al. 2009; Anslow et al. 2008; Klok et al. 2005; Bruan and Hock 2004). The similarity of melt and energy balance conditions at Kbase to other alpine and high latitude regions means that the results displayed here have potential applicability to regions beyond Kbase.

One of the most important results from this study was the amount of densification that the snow experienced over only 16 days. The snow density increase through meltwater addition implies that despite surface lowering and snowmelt, mass is not necessarily lost from the glacier. Refrozen water in the snowpack neither runs off the glacier nor contributes to sea level rise. This means that airborne altimetry studies, such as those used by Arendt et al. (2002) and Berthier et al. (2010), cannot reliably gauge changes in mass balance at sites similar to Kbase on sub-annual timescales

without knowledge of the fate of surface meltwater. After a whole year this effect will be reset since we can assume the each winter's snow is deposited on top of a previously densified and melt afflicted summer snowpack.

In terms of water runoff from Kbase, we have found that a maximum of 100 out of 283 kg/m<sup>2</sup> of meltwater leaves the glacier. This is likely an overestimation as some water is undoubtedly retained in deep firn. This fraction of water is important as it leaves the accumulation zone and may lead to negative summer mass balances locally. This observation agrees with the work of Campbell et al. (2011) whose ground penetrating radar work suggests melting well into the accumulation zone of the Kahiltna Glacier to significant depths.

Despite the increase in number and thickness of melt layers with the increase in positive energy balance and temperature, the observations that less than half of the total meltwater is preserved in discrete melt layers and that melt layer stratigraphy is largely stratigraphically controlled call this reconstruction into question. In order to use melt layers as a proxy for temperature, as Kelsey et al (2010) do close to this site, there must be a strong argument that melting is representative of temperature and that melt layers are preserved in the snow for analysis. The first requirement is tentatively met at this site (Winski et al. 2011), but the second is not. With only 1/3 of the melt water generated entering melt layers at Kbase, it is unlikely that melt layers are representative of surface melting, not to mention temperature. Additionally, many melt layers located very close to the surface were re-melted and subsequently deposited lower in the snowpack, confusing any record later to be interpreted. Finally, the evidence from this study shows that melt layers are stratigraphically controlled; their position dictated by transitions in grain size, density and hardness. This means that melt layers found deeper in the snow may not necessarily be older, making the assumption that depth is proportional to time invalid. We, along with Bell et al. (2008), believe that any quantitative relationship between climate and melt layers is difficult to interpret in an environment with this much melt.

Studies attempting to estimate the energy or radiation budget on the surface of a glacier will often estimate albedo based on the age of the snow, or time since the last snowfall (Braun and Hock 2004; Anslow et al. 2008). Our results demonstrate that surface grain size, which is one of the leading determinants of albedo (Warren 1982), is controlled not only by the age of the snow, but also by the existing stratigraphy in place. This means that in areas similar to Kbase, accurate modeling of surface grain size evolution must take into account the snow stratigraphy. As that would be highly difficult for studies attempting to model albedo or energy balance over space, a more realistic alternative would be to acknowledge that over the first 1-4 days of melt following a storm, grain size will increase to the range of 0.5-2 mm after which point grain size will be controlled primarily by the melting and exhumation of buried layers.

Melt has measurable effects on the chemistry of the snow. After only one day of melt we observed large changes in major ions and a consistent redistribution of trace metals in the snow, showing that post-depositional effects must be considered for any interpretation of the chemical records at this site. Changes in chemistry are arguably even greater here than in areas like Svalbard (Iizuka et al. 2002; Virkkunen et al. 2007; Moore et al. 2005; Pohjola et al. 2002) or Baffin Island (Grumet et al. 1998) as only large features in the chemistry signal were preserved after one month of melt.

Perhaps the best illustration of the effect of melt on chemistry is the isotope record. Although isotope stratigraphy is depressed and altered within the upper snow layers, at least compared with profiles from higher sites (Campbell et al. 2011), below the ash layer the record nearly disappears. Winski et al. (2011), Oerlemans et al. (2009) and Takeuchi and Li (2008), have all demonstrated that a decrease in the albedo (in this case from ash) will lead to intensified melting. Presumably, the melt from the Mt. Redoubt ash was sufficient to nearly homogenize the isotopic signal, lending support to the idea that chemical (or at least isotopic) changes in the stratigraphy are proportional to melt quantity.

Regarding ice cores at sites similar to Kbase, our recommendation is similar to that of Koerner (1997) in that they should be interpreted with extreme caution. Our results from the ion and trace metal experiments show that the stratigraphy primarily reflects post-depositional processes rather than atmospheric concentrations. The isotope record is more robust and may be preserved provided that no special events (like ash deposition) occur. These conclusions are more severe

than those of Grumet et al. (1998) and Pohjola et al. (2002), but this is a very different region than those studies.

## CONCLUSIONS

Melt occurs nearly everywhere outside of interior Greenland and Antarctica (Virkkunen et al. 2007) making knowledge of surface melt processes on glaciers crucial in order to understand the cryosphere and climate systems. Our study has added to this understanding through monitoring changes in the snowpack due to melt. The negative mass balances in Central Alaska observed by Josberger et al. (2007) on the Gulkana and Berthier et al. (2010) by remote sensing are put in context by our study. We have shown that densification measurements are needed to measure changes in mass balance on sub-annual timescales in the wet snow zone. We have made an early attempt at relating densification to melting through equation (2), but this relationship must be further tested.

We assessed the use of melt layers as a paleoclimate proxy for temperature at Kbase and although melt layers have successfully been used as a proxy for atmospheric conditions in environments with lower melt rates, we feel that to use this approach in areas similar to Kbase is unwise. We have found, however, that the formation of melt layers seem to be highly related to boundaries in grain size, snow hardness and density.

We evaluated the viability of Kbase as a possible site for ice core paleoclimate work as well. Strong elution of all ionic species and redistribution of trace metals leading to local enrichments of up to 400% in one day lead us to deem this area undesirable for analysis of chemical species in ice cores with the hope of paleo-atmospheric reconstruction. The exception to this is the isotope record, which is considerably more robust, sustaining dilutions of 10-20 ‰  $\delta D$  after one month. Still, volcanic events, such as that seen at 2.25 meters depth at Kbase are enough to destroy the isotopic signal by promoting excessive melting.

There may be potential to use physical properties in ice cores at this site, but that would require a much longer study period, perhaps in the late summer when summer snow stratigraphy first becomes buried. For greater understanding of the processes at work here, more pits are needed for a longer duration and in different regions of space to test the repeatability of these results and assess the impact of the spatial heterogeneity of the snowpack. We encourage future studies focusing on snowpack evolution, glacial paleoclimatology and mass balance to apply and test these findings.

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