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High-frequency observations of melt effects on snowpack stratigraphy, Kahiltna Glacier, Central Alaska Range

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Abstract:

With an increased volume of melt on many of the world's glaciers, study of how meltwater affects the properties of glacial snowpack becomes essential to our understanding of how glaciers will respond to climate change. We address this problem by studying how snow properties changed on sub-daily timescales on the Kahiltna Glacier, Alaska, between May 26 and June 10, 2010. During this period, we dug 1.8-m-deep snow pits twice daily to record the stratigraphy of melt layers, snow hardness, grain size, and density and sampled for hydrogen isotopic composition (δD) on four occasions. From these data, we show that 65% of the melted surface snow infiltrates and refreezes in the snowpack. This leads to a densification of the snow, a 729% increase in volume of melt layers, and a homogenization of isotopic and physical snow properties. From visual and stratigraphic observations, we show that meltwater flow within the snowpack is conducted primarily along lenses and pipes, where melt layers later form, but that more homogeneous capillary-based flow is also important. Finally, we show using isotope ratios that post-depositional alteration is exacerbated with increased melt extent, using the δD profile below a volcanic ash layer as a case study. In the future, similar studies would benefit from this high-frequency monitoring approach to assessing snowpack evolution, as it allows for a greater understanding of short duration processes. New directions for study would include longer-term monitoring efforts over a wider spatial snow pit network. Copyright © 2012 John Wiley & Sons, Ltd.

KEY WORDS melt layers; densification; glaciochemistry; Alaska; snow

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INTRODUCTION

Glacier surface processes are an integral component of the earth system. Not only are glaciers highly sensitive to environmental perturbations, but changes in glacier health have the potential to affect local and global climate, river hydrology, sea level, erosion, and other systems with complex feedbacks. Given the present shifts in climate patterns and atmospheric variables, understanding glacier surface processes becomes crucial in preparing for and predicting upcoming environmental conditions. One of the largest effects of changing atmospheric parameters, such as temperature, for example, is in the duration and intensity of melting on glaciers (Rupper and Roe, 2008, Hoffman *et al.*, 2008; Munro and Marosz-Wantuch, 2009). In Alaska, the location of our study site, conditions are changing rapidly, particularly temperature which has increased by as much as 2.2 °C in the last 50 years (Stafford *et al.*, 2000). This coincides with increased melt rates leading to glacial recession and mass balances which have averaged −0.31 m/year in Central Alaska since 1962 (Arendt *et al.*, 2002; Berthier *et al.*, 2010; Josberger *et al.*, 2007; Luthcke *et al.*, 2008). Given the increasing amounts of meltwater generated on glacier surfaces, the goal of this study is to determine the fate of surface meltwater in the accumulation zone of the Kahiltna Glacier and to quantitatively assess how it alters the near-surface characteristics of the glacier on timescales of days to weeks.

Several previous studies have examined how meltwater alters the snowpack on glaciers. Mathematical modelling and laboratory studies, such as those by Kapil *et al.* (2010), Colbeck (1972), and Pfeffer and Illangasekare, (1990), have shown how melting of snow occurs in a controlled environment and how stratigraphy changes under laboratory conditions. Additional studies by Arthern *et al.* (2010), Bell *et al.* (2008), Virkkunen *et al.* (2007), Grumet *et al.* (1998), and Pfeffer and Humphrey (1996) investigated how snow stratigraphy and density on cold, high-altitude ice caps changed with the introduction of meltwater. Studies have also been done on seasonal snowpacks to assess how water infiltration, snow temperature, and melting affect snow properties in regions experiencing rain and thaw events (Taylor *et al.*, 2001; Conway and Benedict, 1994). Finally, other groups have performed larger scale modelling studies investigating the spatial distribution and magnitude of melt and its effects on the snowpack (Gardner and Sharp, 2009; Marshall and Sharp, 2008).

The focus of this paper is understanding near-surface glacial hydrology and its impact on the physical and

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chemical evolution of the snowpack in a saturated accumulation zone over daily or sub-daily timescales, as less is known about glacier hydrology in the saturated zone at these high temporal resolutions. Summertime melt occurring in polar regions is commonly assumed to refreeze within the same annual layer; by contrast, meltwater in more temperate regions is lost to run-off. However, at many alpine sites, similar to our study site, the division of meltwater refreezing and run-off is more complicated, with a portion of the summertime meltwater refreezing within the snowpack and the remaining lost to run-off. Additionally, high-frequency monitoring of meltwater in saturated snowpacks is necessary to provide insight into process mechanics, which would be difficult to obtain from longer-term studies. Our aim is that this data set and conceptual model of how meltwater is transported and how it alters the snowpack will inform future research in related fields. A meltwater budget in the lower accumulation zone will aid in the calculation of overall glacier run-off and contribution to sea level rise. How snow properties such as albedo, grain size, density, and permeability change with the addition of melt has important implications for feedback in various glacial systems. Also, by establishing how melt layers form in the snowpack, ice core palaeoclimate studies, such as those by Das and Alley (2005) and Kelsey *et al*. (2010), will better be able to reconstruct temperature and energy balance records based on physical stratigraphy, addressing some of the issues raised by Koerner (1997).

STUDY AREA

To assess the impact of surface meltwater on a glacial snowpack, we monitored conditions at Kahiltna Base Camp (Kbase) 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 for a number of reasons. Chemistry data show that the site receives 1 to 2 m w.e. (water equivalent) precipitation per year, much of which is lost to melt in the summer. The large quantity of precipitation and melt allows us to study melt processes at greater resolution than would be possible in areas of lower accumulation. The onset of the melt season also starts in late April to early May, coinciding with our arrival in mid-May, allowing us to observe the transition of the snowpack from less melt-altered to highly melt-altered. Importantly, there is also evidence that the Alaskan climate has been warming disproportionately during the last 50 years and that Alaskan glaciers are responsible for as much as 25% of sea level rise from melting glaciers and ice caps (Arendt *et al*., 2002; Luthcke *et al*., 2008; Berthier *et al*., 2010). Central Alaskan glaciers have experienced mass balances of −0.31 m/year in the last 50 years (Berthier *et al*., 2010).

Our specific study site at Kbase is situated at 2134 m in elevation on the Southeast Fork of the Kahiltna Glacier (Figure 1). As the equilibrium line altitude is near 1800 m in elevation during most years (Campbell *et al*., 2012), Kbase is located within a zone where the summer snowpack is isothermal and saturated. By late May, when this study commenced, melting had already begun with daytime temperatures usually reaching 7–8 °C.

METHODS

Physical parameters

As the goal of this study is to understand how meltwater affects the physical and chemical evolution of the snowpack, we measure both the snow properties and the amount of melt over time. To this end, five ablation stakes were placed about 30 m 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 stake height 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^3) and a scale. Density measurements were taken three times for each reading, averaged, and used in conjunction with the surface lowering to calculate the total amount of ablation $\mathring{A}$ (kg/m^2/h or mm w.e./h):

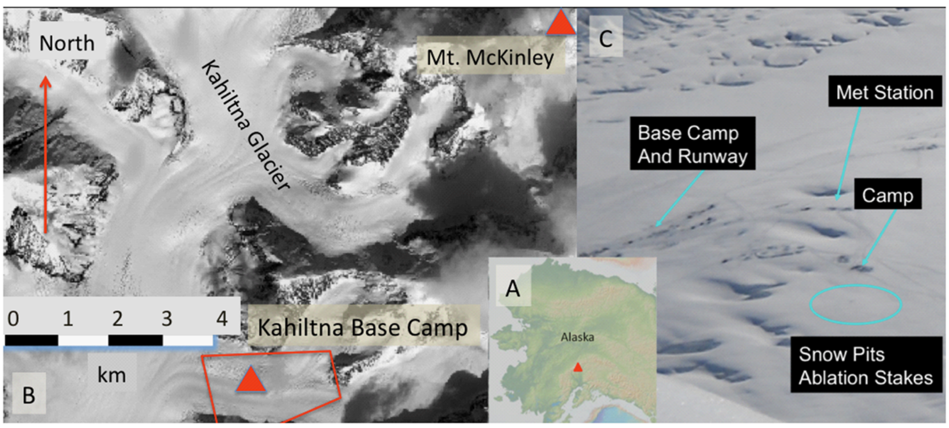

Figure 1. Location map. (A) The position of our site in Central Alaska. (B) The Kahiltna Glacier with the location of Kahiltna Base Camp. (C) Photo of the study site with relevant locations

$$\mathring{A} = \Delta H \frac{\rho}{\Delta t} \tag{1}$$

where ΔH is the average change in snow height referenced to ablation stakes, r is the density of the surface snow in kg/m^3, and Δt is the length of time between measurements. Because each ablation measurement is taken relative to the previous, errors are not propagated over the course of the study. During snow events, it is assumed that melt rate is equal to 0 mm/h, allowing us to measure the surface height immediately after the storm as the new datum. Stakes were read overnight (below freezing conditions) three times, each time yielding no change in surface height, giving us confidence that the change in stake reading each day is primarily driven by ablation, not settling and compaction.

Some ablation, however, is through sublimation and evaporation and does not affect lower snow stratigraphy (Gusain *et al.*, 2009; Stichler *et al.*, 2001; Taylor *et al.*, 2001). 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 (°C), wind speed (m/s), incoming solar radiation (W/m^2), relative humidity (%), and snowpack temperature (°C). 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 whereas 98% enters the lower snowpack as melt.

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), 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-m snow pit was hand dug twice a day between May 26 and June 10, 2010. Pits were dug in a seldom-travelled 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 1–2 m away from a previous pit or disturbed area. We minimized any spatial variation between pits by digging them all within 20 m of one another. All pits had surface slopes of 2°–7° and slope aspects of 270°–305°. Effects of snow redistribution were minimal as we measured similar snowfalls at each ablation stake (surrounding the pits), the snow surface was nearly always either wet or crusted, inhibiting wind transport, and wind speeds at Kbase never exceeded 3.7 m/s, averaging below 1.5 m/s.

We measured snow hardness, grain size, and density in each snow pit. Hardness was measured semi-quantitatively using the method described in the International Classification for Seasonal Snow on the Ground (Colbeck *et al.*, 1990). 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. Melt layers, being any icy lens or layer of large rounded grains held by a matrix of ice, were also recorded at this time.

Chemistry

We assessed the impact of melt on δD over timescales of one day and one month. Upon arriving at Kbase on May 13, 2010, a 3.15-m snow pit was dug in an undisturbed area close to the AWS followed by the extraction of a 6.6-m core from the pit bottom. The pit was scraped with clean plastic scrapers, the stratigraphy was recorded, and continuous snow samples were collected from the pit wall at 15-cm resolution. Samples were collected for δD and placed in Nalgene HDPE 175-ml plastic bottles. Before sampling, bottles for δD were repeatedly soaked and washed in de-ionized water. 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-m snow pit was dug within 3–4 m of the initial pit so as to have the same stratigraphy while not experiencing any thermal alteration from the opening of the initial pit. 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 conducted on the timescale of a day. Early in the morning of June 7, a day forecasted to be warm and sunny, δD samples were collected from that morning's physical properties snow pit. The same sampling and clean procedures were followed as for the two monthly pits. Sampling took place at 10-cm resolution to a depth of 1.3 m. The procedure was repeated that evening after the bulk of the day's melting had taken place. Because these pits were in a different place and sampled at different resolution, no comparisons between the two sets of pit (daily and monthly) are made.

Samples were analysed at the University of Maine Stable Isotope Laboratory. Hydrogen isotopic composition was measured in meltwater from discrete layers in each of the four chemistry pits by gas source isotope ratio mass spectrometry. We used Cr reduction techniques (Morrison *et al.*, 2001) in a Eurovector elemental analyser 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 Standard Mean Ocean Water (SMOW) scale (δD relative to seawater).

RESULTS

Physical properties

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 is estimated to have entered the snowpack during this time. Melt rates were highly variable, ranging from 2.5 mm w.e./h on sunny days to 0 mm w.e./h on cold or stormy days. There was also a difference in the amount of melt at each ablation stake, with the range being from 226 to 297 mm w.e. melted over the study period, depending on the site. Because air temperature was consistently above 0 °C, at least the top 1 m of the snowpack remained isothermal throughout the study period, with the possible exception of the immediate surface after cold nights.

Density of the snowpack recorded over time is shown in Figure 2. Less dense layers tend to have large grains and be softer, although, as opposed to the less directional data in grain size and hardness, there is a very distinct trend of densification over time. At Kbase, the mass of a 1-m^2 area, 1.8 m deep, rose from 956 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, we suggest a linear relationship for densification based on melt:

$$D = 0.56M, \tag{2}$$

where M is melt introduced to the snowpack as calculated by ablation stakes in kg/m^2 and D is mass added to the top 1.8 m of snow in kg/m^2. The site constant for this region is a unitless 0.56, determined by a regression of M and D, and will likely change based on region and time of year. This relation is strong for the 16-day study period ($R^2 = 0.56$) but fails on diurnal timescales, showing that the densification process operates more slowly than the melt process.

One hypothesis for this mass increase is that the snow being removed from the surface is of lower density than firn entering the profile from below (Figure 2). However, barring fresh snow events, both the average surface density and the average bottom density are 600 kg/m^3 ($\pm$ 5 kg/m^3). This suggests that the mass increase originates from the addition of surface meltwater as opposed to differential densities between the top and bottom of the snow pit. We observed that most internal layers maintained their thickness throughout the study and that no settling occurred overnight with reference to the ablation stakes, demonstrating that densification from settling and compaction is minimal. We conclude that the 185 kg of mass added to the snowpack over the study period is principally due to meltwater addition from the surface and that making density comparisons between snow pits over the course of the study period is a valid, if approximate, way of assessing snowpack densification.

The snow hardness data (Figure 3) shows that 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) were intact aside from the fresh surface snow. The banding seen below reflects alternating layers of large grained soft snow and small grained hard snow (Figures 3 and 4).

A plot of grain size over depth and time is shown in Figure 4, which reveals that melt layers most frequently occur at grain size transitions. As the melt season intensifies, crystals in large grained layers tend to shrink whereas 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 (defined as the grain size of the surface layer, from depth = 0 until the depth at which grain size

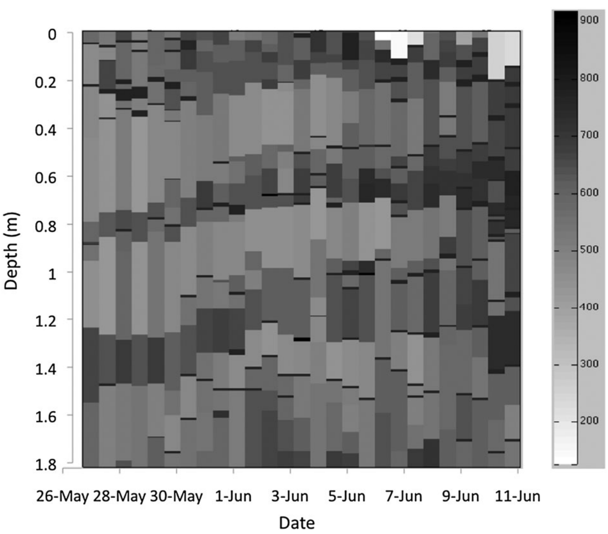

Figure 2. Evolution of snow density of Kbase pits over time

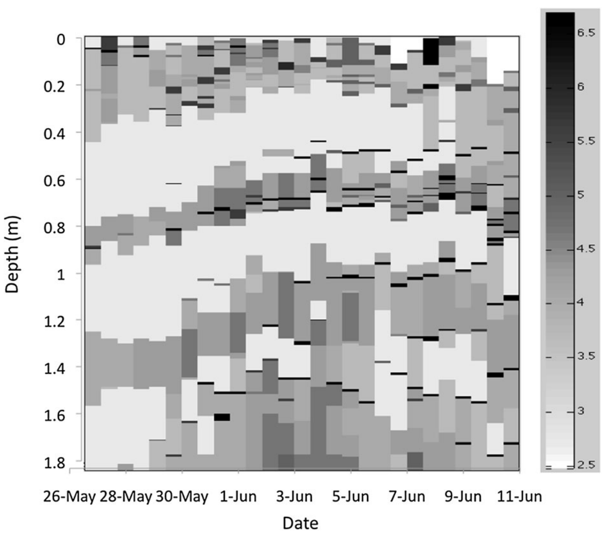

Figure 3. Evolution of snow hardness in Kbase pits over time. The greyscale represents the $\log_{10}$ of the hardness, in Pascals, of the snow as measured by hand

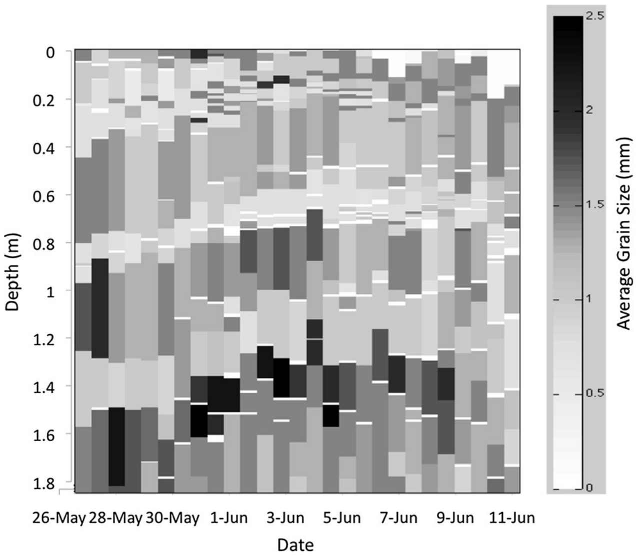

Figure 4. Evolution of snow grain size of Kbase pits over time

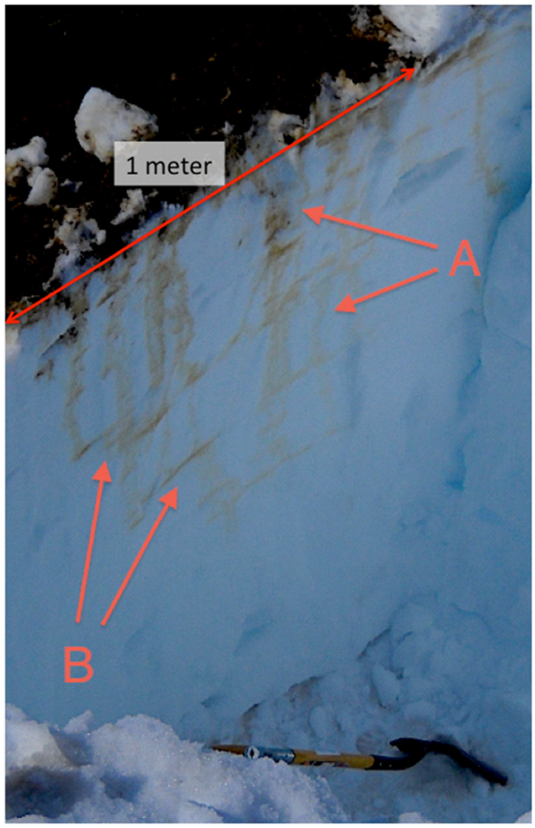

Figure 5. Transport of surface meltwater downwards through vertical pipes (A) and horizontal lenses (B). The width of the soil-covered snow is 1 m

changes) 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 it fluctuated seemingly randomly by a factor of 3. 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 modelling or any study assuming that snow surface properties are atmospherically controlled.

Analysing the patterns contained in these data helps us learn more about the processes at work in the snowpack under melt conditions. By monitoring the movement of individual snow layers, we see that the snow surface moved downwards into the stratigraphy over time as each layer became shallower and, eventually, was exhumed as the surface melted away (Figures 2–4). Over the study period, the surface migrated downwards a total of 375 mm, although the rate was extremely variable. However, in spite of the gradual upward movement of snow layers and changes in their properties, our ability to track layers gives us confidence that we are comparing the same structures among pits over time and that most variation is process driven.

Although similar snow pit profiles are seen in snow pits dug tens of metres and 16 days apart, the mechanisms of meltwater transport through the snowpack have a significant component of spatial heterogeneity. Figure 5 shows a snow pit dug after having set out a layer of soil for 2.5 days (to increase melt through reduced albedo). Soon after the digging of this pit, the stained meltwater channels could be seen infiltrating the snowpack. Although this is not a perfect analogue to a continuous snowpack, this experiment on the snow pit wall shows that meltwater is transported primarily downwards through discreet channels or pipes and horizontally along boundaries in snow properties or impermeable layers.

A degree of spatial discontinuity can also be seen in the snow pit data. Figure 3 shows best how melt layers may be present in one pit and not in the next, even though they were dug hours apart and within 1 or 2 m of each other. Melt layers were occasionally even discontinuous on a single snow pit wall, as suggested by Figure 5. In Brown *et al*. (2011), a similar degree of spatial discontinuity was observed among cores in homologous regions. However, this degree of spatial discontinuity does not necessarily dampen the usefulness of the data, as there is a sufficient volume of snow pit information over a large number of pits such that major and more continuous features, layers, and trends can be seen clearly (Figures 2–4).

There are patterns in the locations where meltwater spreads out horizontally, which is important, as those regions are altered most by the influx of meltwater and are the most likely to form melt layers. On May 26, only three thin melt layers were present in the snow column; by June 10, this increased to 12 melt layers ranging from 5 to 20 mm in thickness and representing a mass increase of 729%. Most melt layers formed at transitions in snow hardness, density, grain size, or a combination of these (Figures 2–4). This may be due to differences in permeability among snow types, leading to water build-up, 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. Observations of Parry *et al*. (2007) are similar to ours in that they did not see a directional preference for melt layer formation at grain size transitions. However, some of the thickest melt layers do seem to have been formed around a 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).

The idea that melt layers form along transitions in both directions of grain size and density is supported by plotting grain size difference between the top and bottom of melt layers versus the density difference on either side of the same melt layers (Figure 6). Two different types of melt layers are highlighted here. The mid-snowpack melt layers are found at moderate grain size and density transitions within the snowpack. The other class of melt layers form after a fresh snow event because 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. A linear regression of the mid-snowpack melt layers 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.

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.

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^2 of snow was lost over 16 days. The upper 1.8 m of snowpack gained 185 kg of mass over the same period. Of the remaining snow lost, 6 kg evaporated or sublimated, leaving 98 kg (or 35%) of meltwater to sink to a depth lower than

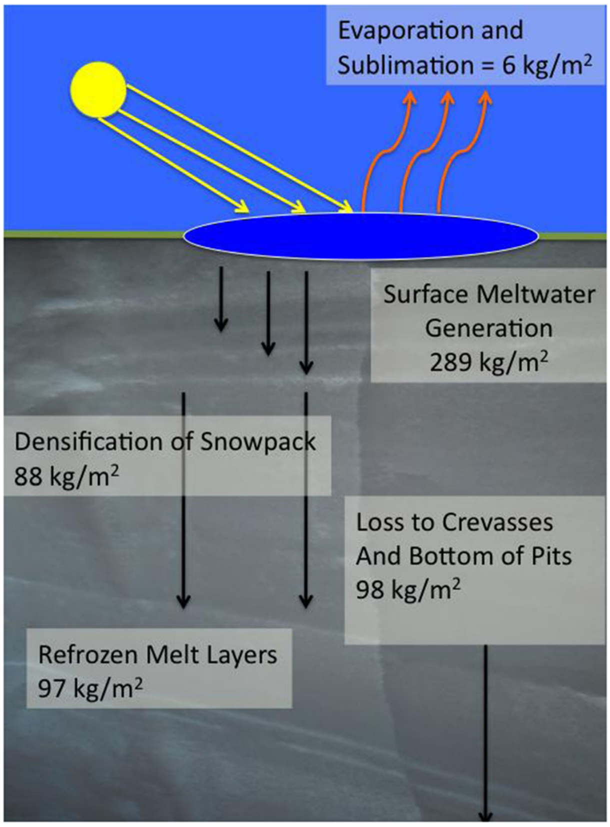

Figure 7. Fate of surface meltwater between May 26 and June 10, 2010, at Kbase. Of the 289 kg/m^2 of ablation at the surface, 6 kg evaporated or sublimated, 97 kg refroze in melt layers, 88 kg uniformly densified the snowpack, and the remaining 98 kg infiltrated to a depth greater than 1.8 m

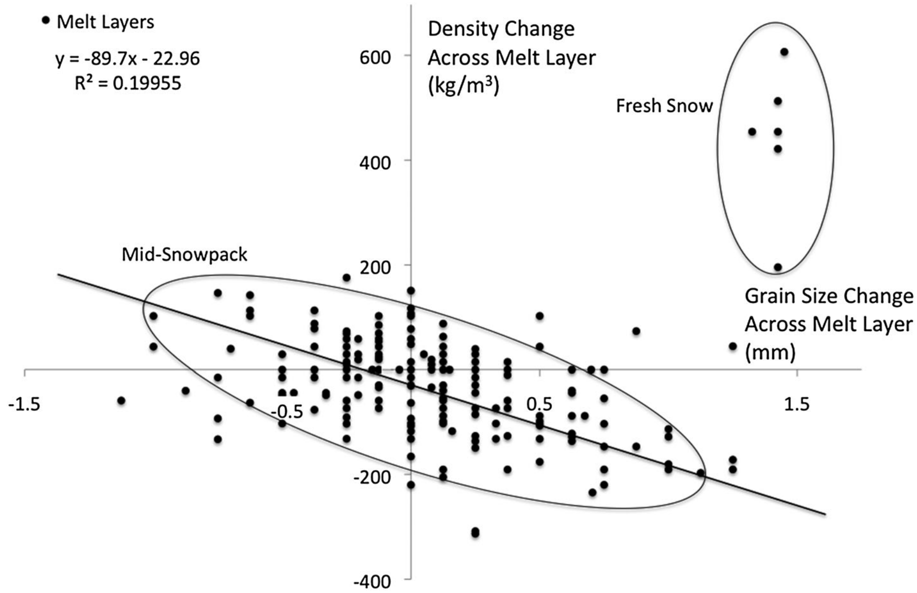

Figure 6. Melt layers recorded in the study, plotted by the difference in grain size from top to bottom of the melt layer versus the difference in density over the same interval. Positive values indicate increasing density and grain size downwards. The two principle types of melt layers are shown on the plot as is the regression line for the mid-snowpack melt layers

1.8 m (Figure 7). We assume that any horizontal flow into an area will equal that out of it because an important assumption we make is that all adjacent snow patches are experiencing similar conditions.

Of the 185 kg of the meltwater that remained in the upper snow column, about 97 kg or 52% 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 88 kg of water went into increasing the density of softer layers in the snowpack, demonstrating the importance of continuous capillary-driven flow as opposed to solely pipe and lens facilitated meltwater transport. These figures represent our best estimates given that they are based on highly spatially and temporally variable data that were averaged to produce this budget.

Chemical properties

Stable isotope data provide a good illustration of how and where melt alters snowpack stratigraphy. We sampled δD in the morning and evening of June 7, at 10-cm intervals in the top 1.3 m of a snow pit (Figure 8). A total of 19 mm of melt at the surface occurred between the sampling times. The average δD 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.

The persistence of the δD profile over the course of one day not only is important in terms of describing the behaviour of isotope ratios but also shows that we can make comparisons between the two pits in a meaningful way. If the curves in Figure 8 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.

We compared δD 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 sampling dates. To match the sampling depths in the two pits, we used our knowledge of surface lowering (37 cm), the location of a visible ash layer, and major peaks in the δD profiles to linearly scale the depth intervals in the June 10 pit such that they would correspond to depths in the May 13 pit.

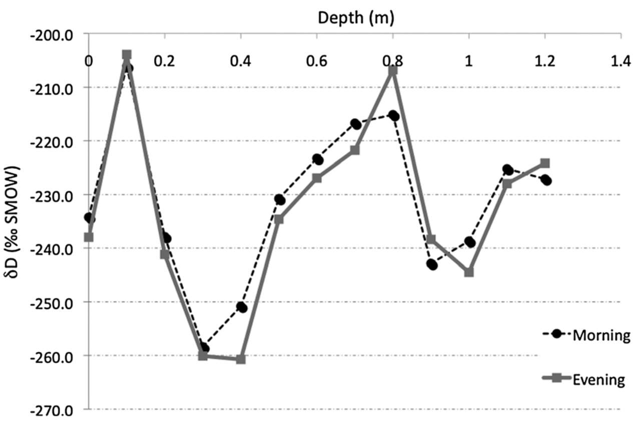

Figure 8. Plots of δD in adjacent snow pits in the morning and evening of June 7, 2010, at Kbase

Similar to the physical snow properties, the isotope data show that the chemical stratigraphy becomes more homogenized with the increase in meltwater addition (Figure 9). 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 δD across the profile remains stable. After a month, the surface has higher values of δD, whereas lower depths have lower δD. Our hypothesis is that during melting and sublimation, ^{1}H is preferentially converted to liquid and vapour whereas more ^{2}H is left behind in the upper snowpack. The meltwater, now depleted in ^{2}H, percolates downwards, depositing more ^{1}H and less ^{2}H 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 such as that described in Cuffey and Steig (1998) is likely responsible for some of the change.

At 2.25–2.3 m depth, there is a dark visible ash layer, which was deposited in the spring of 2009 from the Redoubt Volcano. Below the ash layer is a region in which the isotopic profile has nearly uniform values of δD at −184 ± 2‰. We believe 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 the reduced albedo, was sufficient to homogenize the original

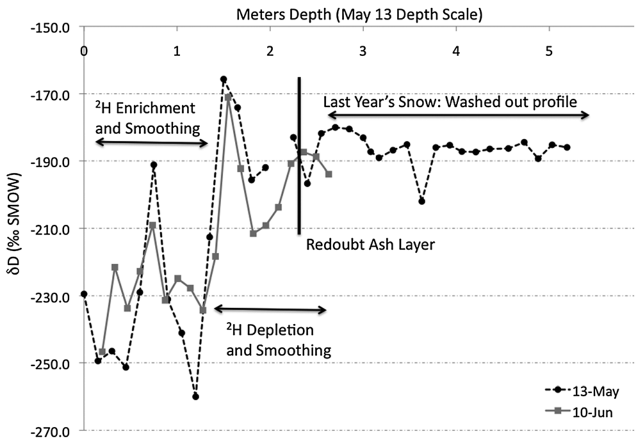

Figure 9. Changes in deuterium profile at Kbase after one month. Peaks were subdued with slight increases in δD in the uppermost metre and δD decreases below that. Below the ash layer, the δD profile was almost completely homogenized, owing to the increase in melt from the lowered albedo at this time

isotopic profile, supporting our findings that although melt layer stratigraphy becomes more complex with the addition of melt, both physical and chemical properties become more homogenized. It is difficult to assess if the isotopic profile would be homogenized in a normal summer, as we do not have a sample of snow that has undergone summer melt but with no ash layer. However, we use this as an illustration of how greater degrees of melt lead to greater alteration and, in this case, homogenization of the snow profile.

DISCUSSION

An overall synthesis of the data is presented here as the following conceptual model of how meltwater is transported and deposited in the snowpack: A surplus of energy at the snow surface leads to a change in phase of the surface snow, almost all of which is a solid to liquid melting transition. When melting occurs, surface grain size coarsens appreciably and is eventually melted away, exposing lower stratigraphy. Once liquid, meltwater is pulled down into the snow through gravity and capillary action, as seen in Colbeck (1972) and Kapil *et al.* (2010). Meltwater moves through the snow primarily in lenses and vertical pipes, controlled by micro- and mesoscale spatial variations in the snow structure, in agreement with work done by Humphrey *et al.* (2012), Campbell *et al.* (2005), and Parry *et al.* (2007). Upon reaching impermeable structures previously in place, such as buried wind crusts, or older melt layers, the meltwater spreads out along these planes. Sometimes the water refreezes here, resulting in thick ice layers and a release of latent heat, maintaining the snow temperature at 0 °C. Also common is a growth in grain size along these layers from the water refreezing onto grains at these stagnant areas. Water spreads out along these regions until it finds a weak area, below which vertical piping occurs, or the impermeable layer is softened and melted enough for water to penetrate. Both cases have been observed in our study.

A numerical value for the volume of water transported by pipes, channels, and lenses versus that transported more uniformly by capillaries and matrix flow cannot be calculated from our data. However, we do know that for meltwater remaining in the upper snowpack, the ratio is almost 50 : 50 between mass that is refrozen in discrete layers as opposed to that more uniformly refrozen in the snowpack. In terms of melt layer mass as a percentage of total snow with a column, melt layers at Kbase increased from less than 1% on May 26 to 10% by June 10. This is already more than an entire season's worth of melt layers in 4 of the 5 years studied at Kahiltna Pass (Kelsey *et al.*, 2010). For our study, 35% of meltwater infiltrates deeper than 1.8 m, the depth of our pits. This is a great deal more retention than what was observed on the Taku Glacier in Coastal Alaska, where 79% to 90% of meltwater was not retained (Pelto and Miller, 1990), but as the snow becomes saturated, the 35% of meltwater infiltrating deeper will increase. It is likely that this meltwater refreezes in the deeper snowpack, although we lack the data to confirm whether or not any meltwater from our study site contributes to glacial run-off.

Perhaps the best illustration of the effects of melt on stratigraphy is the snowpack stable isotope profile. Although isotope stratigraphy is altered within the upper snow layers, at least compared with profiles from higher sites (Campbell *et al.*, 2012), below the ash layer, the record of variability nearly disappears. Oerlemans (2009) and Takeuchi and Li (2008) have demonstrated that a decrease in the albedo (in this case from volcanic ash) will lead to intensified melting. Presumably, the melt induced by the Mt. Redoubt ash was sufficient to nearly homogenize the isotopic signal, lending support to the idea that stratigraphic (or at least isotopic) changes in snow properties are proportional to melt quantity. This finding is supported by the work of Kelsey *et al.* (2010) at a nearby site, who found stratigraphic alteration (melt layer formation) to be proportional to temperature-induced melting.

Our results confirm the findings of Brown *et al.* (2011), Kelsey *et al.* (2010), Iizuka *et al.* (2002), and Grumet *et al.* (1998) and other similar studies in that they show that post depositional snowpack alteration increases with intensity of melting. The mechanisms of vertical piping, lensing along impermeable boundaries, and the importance of capillary action are also demonstrated on the Kahiltna Glacier. Our results are also roughly in agreement with those of Brown *et al.* (2011) in that both studies document a similar amount of spatial variation on the scale of roughly 15 m. Many other papers using melt layers as a palaeoclimate indicator have been completed primarily in regions such as Antarctica, Svalbard, or Greenland, where the snowpack is colder and melt refreezes closer to the surface (Das and Alley, 2005; Herron *et al.*, 1981; Iizuka *et al.*, 2002). Given our observations that melt layer stratigraphy on daily to weekly timescales reflects previous structures and flow patterns as much as climate, we would recommend that palaeoclimate studies either avoid areas with this degree of melt or use a melt percent indicator similar to that used in Koerner (1977) as opposed to attempting interpretation of individual events.

Future directions for study would include arriving earlier in the season to assess a winter snowpack before any alteration has occurred and subsequently monitor snow conditions through the onset of the melt season into late summer. Our observations of early season melt and snowpack evolution represent only a small slice of the changes likely to occur over an entire summer. It is likely that further chemical alteration and stratigraphic homogenization will continue over the summer and that an extension of our study later into the season would show the manner in which these processes occur. However, given our study window of 16 days, we assert that the period in late May to early June is among the most important to understand, as the increase in melt volume is greatest at this point in the year. A greater quantification of snow temperature and subsurface energy balance would also strengthen the conceptual model we present, as would a rigorous spatial analysis of the heterogeneity of snow properties over depth. To be able to confidently apply the findings of this study to other sites, it is essential for studies such as this to be conducted with a wider spatial array of snow pits. Fortunately, our study site was uniform and compact enough that the snow properties seen between pits match well and tell a coherent story of densification, melt layer formation, and homogenization with increased melting.

In regard to run-off calculations from glaciers, we have constrained the amount of run-off generated per metre at Kbase during our study period to between 0 and 98 kg. This is obviously not precise enough for quantitative estimates of run-off contributions from accumulation areas, but it shows that these areas should not be ignored in run-off and sea-level projections, nor should it be assumed that all melt is infiltrating and reaching the bed in these situations. The implications of this second point are that remotely derived estimates of surface lowering as a proxy for glacial mass loss are not at all appropriate in such a region as Kbase on such a timescale due to the melt-induced densification of the upper snowpack. In the future, deeper snow pits, buried moisture sensors, and a more complex analysis could help better assess the run-off contribution of highly melted accumulation areas.

CONCLUSIONS

Melt occurs nearly everywhere outside of interior Greenland and Antarctica (Virkkunen *et al*., 2007), making knowledge of surface melt processes on glaciers crucial to understanding the cryosphere and climate systems. Our study has added to this understanding, as it is one of the most consistent short-term monitoring efforts on snowpack evolution providing detail on how and where changes occur in a previously unstudied region. We show the processes by which summer melt takes place and that regions low in the accumulation zone undergo significant surface melting. This contextualizes the negative mass balances in Central Alaska observed by Josberger *et al*. (2007) on the Gulkana and by Berthier *et al*. (2010) with remote sensing. We have shown that densification measurements are needed to measure changes in mass balance on sub-annual timescales in the accumulation zone. We have made an early attempt at relating densification to melting through Equation (2), but this relationship must be further tested.

We conclude that the addition of more meltwater in a snowpack has an unambiguously positive relationship with the degree of post-depositional alteration. This is perhaps best illustrated in the isotope record, which was hardly altered after one day of melt, significantly smoothed after one month, and almost completely washed out after the albedo induced melt increase associated with the 2009 Mt. Redoubt eruption.

We produced a record of change in quantitatively measured physical snow properties at high temporal resolution, allowing us to further understand snowpack processes. Through these observations, we propose a conceptual model of meltwater transport within the snowpack, melt layer formation, and alteration of physical and chemical snow properties in relation to melt. The finding that meltwater is transported both uniformly and through pipes and lenses can be tested at other sites. Similarly, our findings that 35% of meltwater infiltrates deeper than 1.8 m whereas 34% is refrozen in melt layers and 31% densifies pre-existing snow layers more homogenously should be tested in other regions, ideally to form a relation of how all these numbers change with the amount of summer melt.

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