Seasonal trends in the stable isotopic composition of snow and meltwater runoff in a subarctic catchment at Okstindan, Norway

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Abstract Seasonal variations in the stable isotopic composition of snow and meltwater were investigated in a sub-arctic, mountainous, but non-glacial, catchment at Okstindan in northern Norway based on analyses of $\delta^{18}O$ and δD. Samples were collected during four field periods (August 1998; April 1999; June 1999 and August 1999) at three sites lying on an altitudinal transect (740–970 m a.s.l.). Snowpack data display an increase in the mean values of $\delta^{18}O$ (increasing from a mean value of -13.51 to -11.49‰ between April and August), as well as a decrease in variability through the melt period. Comparison with a regional meteoric water line indicates that the slope of the $\delta^{18}O-\delta D$ line for the snowpacks decreases over the same period, dropping from 7.49 to approximately 6.2.This change points to the role of evaporation in snowpack ablation and is confirmed by the vertical profile of deuterium excess. Snowpack seepage data, although limited, also suggest reduced values of δD, as might be associated with local evaporation during meltwater generation. In general, meltwaters were depleted in $\delta^{18}O$ relative to the source snowpack at the peak of the melt (June), but later in the year (August) the difference between the two was not statistically significant. The diurnal pattern of isotopic composition indicates that the most depleted meltwaters coincide with the peak in temperature and, hence, meltwater production.

Keywords Evaporative losses; late-lying snow; snow metamorphism; water isotopes

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Introduction

The distributions of naturally occurring environmental isotopes in the hydrosphere are often used to constrain hydrological, hydrogeochemical and biological problems, such as water budgets and routing in catchments (reviewed in Sklash (1990) and Kendall *et al.* (1995)), sources of groundwater recharge (e.g. Saxena 1984; Maule *et al.* 1994; Abbott *et al.* 2000) and solutes (e.g. Erel *et al.* 1990; Miller *et al.* 1993; Karim and Veizer 2000), as well as biogeochemical cycling (e.g. Nadelhoffer *et al.* 1988; Aravena *et al.* 1992; Durka *et al.* 1994; Velinsky and Fogel 1999). For strictly hydrological applications, the heavy isotopes of hydrogen and oxygen are the most useful, due both to their relative abundance in a given water sample as well as their insensitivity to geochemical and biological processes within catchments. Whilst the stable isotopes of oxygen and hydrogen have been widely applied in catchment studies in a range of environments, including glaciated catchments (Theakstone 1988; Theakstone and Knudsen 1996a) and those with significant seasonal snow covers (e.g. Obradovic and Sklash 1986; Cooper *et al.* 1993; Taylor *et al.* 2001), the relationship between isotopic variations in snow and the meltwater produced from it are, in general, still poorly characterised (Unnikrishna *et al.* 2002). Of particular importance is the evolution of the isotopic composition of a snow cover through the melt season as it is subjected to melting, sublimation and refreezing, as well as to evaporation at the snowpack surface and vapour transport processes within and beneath the snowpack. Of potentially equal significance is the fractionation that occurs during partial melting of a snow cover,

producing a meltwater that is depleted in the heavier isotopes of oxygen and hydrogen, relative to the snowmelt source. Both of these factors contribute to temporal variations in the isotopic composition of meltwater runoff that undermine the use of simple mixing models for estimating snowmelt *vs*. rainfall, soil water or glacial meltwater contributions to runoff which do not take these trends into account.

Early work which used water isotopes to analyse snowmelt processes in catchments was largely concerned with characterising the relative contribution of snowmelt to runoff (e.g. Dinçer *et al.* 1970; Sklash and Farvolden 1979; Rodhe 1981; Hooper and Shoemaker 1986) as a basis for hydrograph separation. More recently, some of the motivation for catchment-scale analysis of snow meltwater pathways has derived from concern over acidification during spring snowmelt (e.g. Bottomley *et al.* 1986; Maule and Stein 1990; Neal *et al.* 1997; Soulsby *et al.* 2000), as well as from the need to estimate glacial ablation based on runoff with a significant seasonal snowmelt component (e.g. Theakstone and Knudsen 1989, 1996b). In addition to catchment-scale studies, temporal changes in the isotopic stratigraphy of a snowpack have also received attention (Moser and Stichler 1975; Martinec *et al.* 1977; Stichler 1981, 1987; Rodhe 1987 as cited in Cooper 1998; Friedman *et al.* 1991; Sommerfield *et al.* 1991; Raben and Theakstone 1994, 1998; Sturm and Benson 1997; Taylor *et al.* 2001; Unnikrishna *et al.* 2002) with results indicating that some isotopic homogenisation occurs within a snowpack during metamorphism and depth hoar formation prior to melting and that the snowpack becomes progressively enriched in heavier isotopes through the melt season. At the snowpack surface, the isotopic content may be enriched by evaporation, whereas vapour transport at the base of the snowpack can have a similar effect, particularly where the thermal gradient is large (Moser and Stichler 1975; Friedman *et al.* 1991; Sturm and Benson 1997). The relationship between the isotopic composition of the snowpack and of the meltwater derived from it (prior to mixing with other water sources) has received somewhat less attention, and several hydrological studies have assumed that the two compositions should be identical and have, accordingly, used snow cores to assess snow meltwater contributions to runoff. However, as demonstrated by Herrmann *et al.* (1981) and more recently by others (Taylor *et al.* 2001; Unnikrishna *et al.* 2002) the correspondence between snowpack and snow meltwater "signatures" varies significantly, but also systematically, through the melt season. As might be anticipated, the meltwater is generally depleted relative to the snow, although variability in meltwater flux rates and in the portion of a snowpack generating melt preclude a simple relationship between the two.

This study examines seasonal trends in the stable isotopes of oxygen and hydrogen in snowpacks and in the snow meltwater derived from them in a subarctic catchment at Okstindan in Norway. The analysis is based on samples collected at three times during a single "snow year": early April, representing a largely "pre-melt" snowpack; early to mid-June when the seasonal "peak" melt is well under way; and early August, after the bulk of melting has occurred and only persistent "residual" seasonal snow remains in the area. It was possible, therefore, to estimate the magnitude of the change in the isotopic "signature" of a snowpack through the melt season and to examine modifications to this signature along a flow path under peak melt (June) and residual melt (August) conditions. By working with both heavy oxygen and hydrogen, it was also possible to consider the potential contribution of evaporation and vapour transport to observed patterns of relative isotopic enrichment or depletion.

The study area

Snow and water samples were collected in the vicinity of Okstindtjønna, an upland lake (752 m a.s.l.) located on the eastern side of the mountainous area of Okstindan, subarctic Norway (66°02′ North, 14°24′ East) close to the border with Sweden (Figure 1). Previous

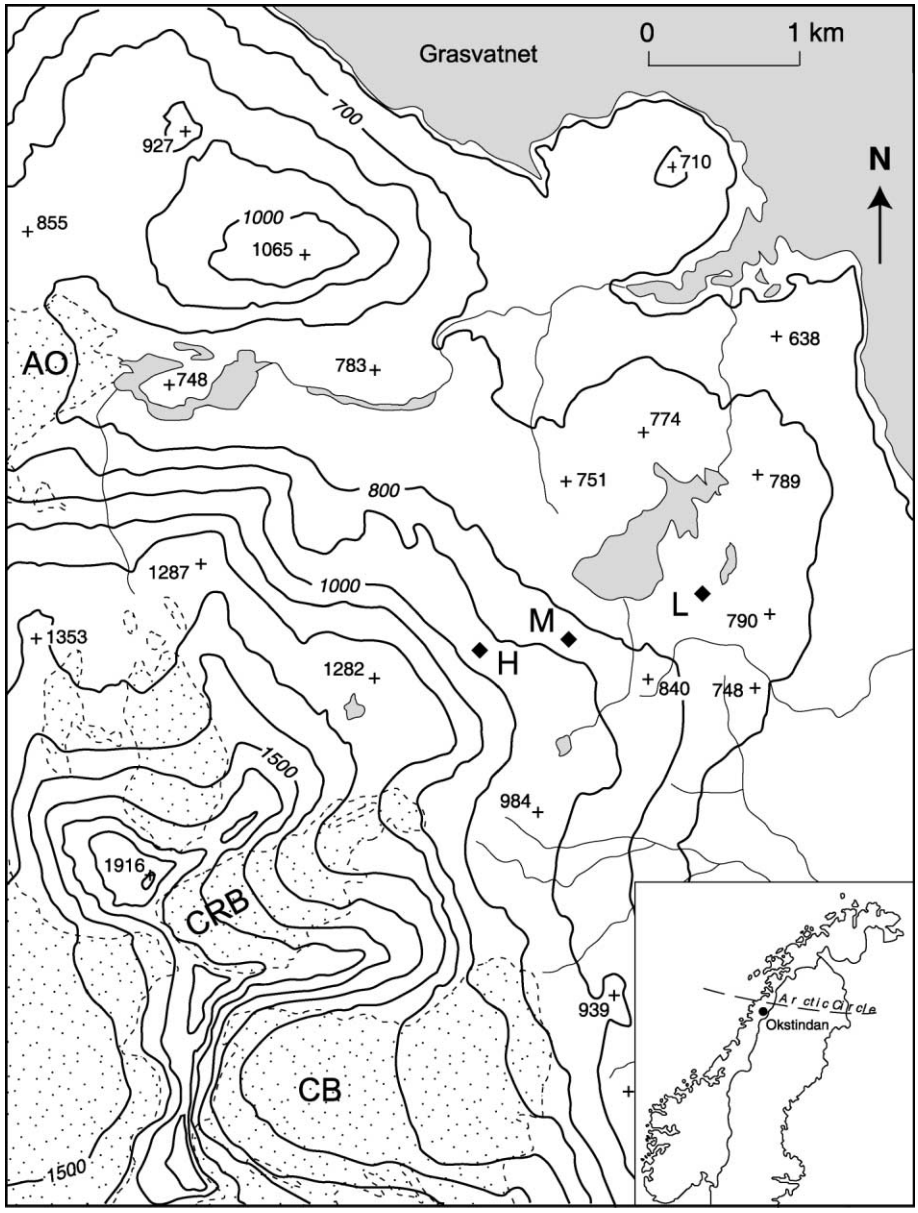

Figure 1 The Okstindtjønna study area and location of Okstindan within Norway. Snow sample sites (L - Low, M - Middle and H - High). Glaciers: AO - Austre Okstindbreen, CRB - Charles Rabot-breen and CB - Corneliussen-breen

studies in this area have considered the glaciology and geomorphology of the glacial systems (e.g. Knudsen and Theakstone 1988), solifluction processes (e.g. Elliott and Worsley 1999) and soil development (e.g. Ellis 1980a).

No detailed climatic records exist for the site: however, previous workers have estimated a mean annual air temperature of around -0.5°C and an average annual precipitation of approximately 1500 mm/yr (Ellis 1980b). Precipitation is mostly in the form of snow which falls, accumulates and is redistributed by wind in the period September/October to April/May. In the winter, cold easterly air can dominate for considerable periods.

The summer climate is cool and wet and driven primarily by moist, westerly airstreams. The ground temperature regime is affected greatly by snow cover. In areas with thick and persistent snow cover, soils suffer only a seasonal freeze/thaw cycle and are not affected by diurnal freeze/thaw cycles in the autumn or spring (Harris 1974).

The local area is dominated by the mountains of Oksskolten (1916 m) and Okshornet (1901 m) to the south-west and by the Okstindan plateau ice cap, an outlet of which, Austre Okstindbreen, lies to the west-north-west. The smaller isolated cirque glaciers of Corneliussen-breen and Charles Rabot-breen lie to the south-west (Figure 1). Streams drain the margins of these temperate glaciers and, due to their heavy winter snow cover, produce numerous snow meltwater streams in the late spring and early to mid-summer. The area has an extensive network of small and large lakes which are maintained by direct surface runoff from snowmelt and summer precipitation, as well as by groundwater discharge.

The soil cover, where present, is developed either directly on bedrock (locally comprising a siliceous mica-schist) or on a substrate of glacial till, which occurs in scattered patches. The soil varies in thickness from 1 or 2 cm to over 0.5 m, with large variations occurring over short distances. The main soil types include Brown Soils, Humic Regosols, Podzols and Regosols, the latter having an association with the location of late-lying snow patches (Ellis 1980b, c).

The area around Okstindtjønna is botanically rich and lies just above the transition between the subalpine birchwood zone to the Mid- to High-Alpine Zone (Holyoak 1982). All of the sample sites lie above the tree-line (*Betula pubescens* forest) and here the vegetation cover is variable, with several species of dwarf shrubs (including *Salix herbacea* and *Salix lapponum*) growing in the hollows and wetter areas. In drier areas, where bedrock is exposed, there is an incomplete plant cover of bryophytes, lichens and small flowering plants such as *Loiseleuria procumbens* and *Diapensia lapponica*.

The snow cover which accumulates in the area around Okstindtjønna throughout the winter is considerable, and redistribution of snow by the wind can create a winter snowpack with a thickness exceeding 10 m in some localities. In other areas, such as on ridges and summits, the snow cover is almost totally deflated by the wind. As with all mountainous areas, the altitude, topography and aspect of a site determines whether a late-lying, semi-perennial or perennial snowpack will develop (McKay and Gray 1981; McClung and Schaerer 1993; Ferguson 1999). Snowpack location is clearly not simply related to the depth of snow accumulation, but rather has a closer connection to the level of protection that sites have from direct solar insolation, wind and higher air temperatures. Hence, large depressions aligned along north-facing slopes at altitude are the most favourable for the persistence of late-lying snow and the development of perennial snowpacks, provided that sufficient snow accumulates there in the winter.

Field methodology

Prior to the initiation of the 1999 study reported here, three sampling sites were identified in the vicinity of Okstindtjønna during a field visit in August 1998. The sites selected lie on a rising transect from 740–970 m a.s.l. and have either a late-lying snow patch which always melts out by August (the Low site in Figure 1), or a more persistent semi-perennial or perennial snowpatch (the Middle and High sites in Figure 1, respectively). A limited number of snow and meltwater samples were collected from these sites in August 1998, representing "residual" conditions prior to the onset of the next snow year.

In 1999, three periods of fieldwork enabled samples to be collected from the pre-melt snow (April), the peak melt phase (June) and the residual melt phase (August) from the three sites. Snow samples were obtained at various depths from the surface of the snowpack through the construction of snow pits. One or two pits were excavated at each site during

each field period, for a total of 14 snow pits for 1999. The sides of the pits were cleaned using an aluminium snow shovel and the physical stratigraphy of the snow was logged at centimetre intervals recording hardness (on a six-point scale from very soft to solid), moisture content (on a five-point scale from dry to very wet), grain size (in mm) and grain shape (using a six-point classification based on a simplification of the scheme presented in Colbeck *et al.* (1990)). Snow samples were cut out of the back face of the pits at uniform sampling intervals of 400 mm (the average sample volume being ~ 1 litre) and placed in nested, sealed polyethylene (PE) bags (see Fischer *et al.* 1995). These were then allowed to melt under ambient conditions and the resulting meltwater was sampled directly into 30 ml high-density polyethylene (HDPE) screw-cap sample bottles, after agitation of the sample bag to ensure homogenisation. Snowmelt seepage was collected by allowing water dripping from the lower edge of the snowpack (for example, at the base of snow pits or from the downslope edge of snowpacks) to fall directly into the sample bottles. Meltwater runoff in streams downstream from the edge of the snowpack was collected by allowing the water to flow directly into the sample bottles. Soil seepage water was collected by digging shallow soil pits and allowing the sub-surface seepage to flow directly into a sample bottle. Once water samples were secured in the bottles, the caps were sealed with electrical insulating tape and the bottles packaged in sealed PE bags for return to the laboratories.

At one site (the Middle site), a four-hourly sampling programme was conducted over 28 hours during the June field session. Samples were gathered at points along a transect from the edge of the actively melting snowpack of the Middle site, through the shallow soils immediately downstream of the snowpack, and into and through a meltwater stream draining ultimately into Okstindtjønna (Figure 2).

Stable isotope analyses

Deuterium and ^{18}O concentrations were obtained through isotope ratio mass spectrometry, with the hydrogen analyses conducted by The University of Reading, UK and the oxygen

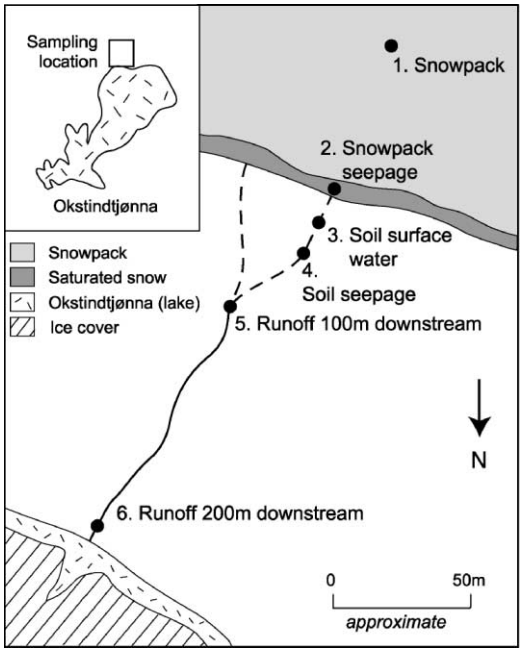

Figure 2 Sampling locations along a meltwater pathway from the snow pit at the Middle site in June to the point of meltwater discharge into Okstindtjønna

analyses conducted by the British Geological Survey at Wallingford, UK. The hydrogen analyses were conducted using the zinc reagent reaction technique and a dual-inlet stable isotope ratio analyser mass spectrometer (VG SIRA 10 – Series II). The oxygen analyses were conducted using the CO_2 equilibriation technique and a dual-inlet isotope ratio mass spectrometer (VG Optima).

Results of the analyses are expressed as delta values, δD (the enrichment of ^{2}H or D with respect to ^{1}H) and δ^{18}O (the enrichment of ^{18}O with respect to ^{16}O), in parts per thousand (‰), relative to a standard, V-SMOW. Analytical precision is generally in the range of 0.05–0.2‰ for δ^{18}O and 0.2–2.0‰ for δD (Kendall *et al.* 1995).

The isotopic composition of precipitation at a given location is determined by condensation temperature (and, therefore, latitude, altitude and season) and by the degree to which the water vapour in the air mass has previously produced precipitation during its movement from an oceanic source. Winter precipitation tends to be more depleted in the heavy isotopes, although inter-storm variations in a given season can be significant (Kendall *et al.* 1995). In the absence of evaporation, the relationship between δ^{18}O and δD is given approximately by the Global Meteoric Water Line (GMWL), i.e. $\delta D = 8\delta^{18}O + 10$ (Craig 1961), although, where available, a Local Meteoric Water Line (LMWL) is preferred. For this study, as a LMWL was not available for the Okstindan area, a "regional" meteoric water line (hereafter referred to as RMWL) was established using the Global Network for Isotopes in Precipitation (GNIP) archive (International Atomic Energy Agency, Vienna Austria: http://isohis.iaea.org), based on the four most adjacent stations (at Ny Ålesund and Isfjord Radio, Svalbard, Norway, Lista, Norway and Naimakka, Sweden). This allows comparison of an approximate regional δ^{18}O–δD line with those of the snow and water samples collected at Okstindan.

When evaporation occurs, the remaining water is preferentially enriched in ^{18}O relative to D, resulting in a δ^{18}O–δD of lower slope than that of the original water. Although this isotopic shift is well characterised for equilibrium fractionation occurring during the evaporation of free water (resulting in a line of slope = 5), non-equilibrium conditions associated with evaporation in porous materials can produce slopes as low as 2 (Dinçer *et al.* 1974; Allison 1982; Allison *et al.* 1983). This reduction in the slope of the δ^{18}O–δD line has also been observed during snowpack metamorphism, as reported by Friedman *et al.* (1991) and Sommerfield *et al.* (1991), and snowpack melting. Cooper *et al.* (1993), in particular, noted that, during melting of a snowpack in the Alaskan arctic, both the snowpack and the streamflow derived from it had slopes of between 6 and 7 (in contrast with the slope of the LMWL of 8.28) and attributed these lower slopes to evaporative isotopic enrichment during snowmelt.

Results

A total of 271 (226 original and 45 replicate) snow and water samples were collected during four field visits (August 1998, April 1999, June 1999 and August 1999). All samples were analysed for δ^{18}O (‰) and about half were analysed for δD (‰). Field and laboratory replicate samples indicate an average error of 0.19 and 0.125‰, respectively, for δ^{18}O. The average error in the analysis of δD was 1.92‰. The results of the oxygen analyses are summarised in Table 1 and indicate that values of δ^{18}O range from −22.6 to −8.34‰. Mean values vary considerably by sample type and time of sampling and range from relatively depleted values of −14.47 and −15.17 for June snowpack seepage and soil seepage waters, respectively, to −9.45 for late-lying snow (August 1998).

Table 1 Results of $\delta^{18}O$ analyses (‰), grouped by sample type and sampling month

Sample type	Minimum	Maximum	Mean	Standard deviation	No. of samples
Snowpack (August 1998)	−10.44	−8.34	−9.45	0.92	6
Snowpack (April 1999)	−22.60	−8.42	−13.51	3.04	30
Snowpack (June 1999)	−16.87	−10.37	−13.26	1.42	68
Snowpack (August 1999)	−14.88	−10.08	−11.49	1.27	15
Seepage (June 1999)	−16.89	−8.85	−14.47	1.85	15
Seepage (August 1999)	−15.46	−10.68	−11.97	1.60	7
Runoff (June 1999)	−16.62	−9.33	−14.28	1.44	30
Runoff (August 1999)	−12.62	−10.37	−11.28	0.72	12
Soil water (June 1999)	−21.87	−12.0	−15.17	2.09	26
Rain (1999)	−10.54	−8.50	−9.77	0.68	6

Isotopic and physical evolution of the snowpack

The 1999 snowpack samples indicate both an increase in the mean value of $\delta^{18}O$ from a pre-melt value of −13.51 to −11.49‰ for late-lying snow, as well as a decrease in variability with the range decreasing from 14.18 to 4.8 and the standard deviation decreasing from 3.04 to 1.27 over that time. However, Figure 3 illustrates the mean value and standard deviation at each of the three sampling sites and indicates that the extent of these trends does vary somewhat between sites. The Middle site (840 m a.s.l.) exhibits a significant increase in mean $\delta^{18}O$ values between April (pre-melt) and June (peak melt), particularly when compared with the Low site (740 m a.s.l.) and High (970 m a.s.l.) sites, which both appear to maintain a constant mean value over that time period. Full excavation of the snowpack was not feasible at the Middle and High sites during April, however, so the mean values only reflect a partial excavation of the upper 2–2.5 m at those sites. Snowpacks at all three sampling sites, though, do exhibit a pronounced decrease in variability through the sampling season, indicating the progressive isotopic homogenisation of the snowpack. These general trends are consistent with the isotopic profiles in individual pits for successive sampling periods, eight of which are illustrated in Figure 4. The progressive enrichment and homogenisation at the Middle site is well represented in Figure 4, whilst the overall homogenisation with little isotopic shift observed at the High and Low sites is also apparent.

The isotopic evolution of the snowpacks may also be considered in terms of the relative fractionation of δD against $\delta^{18}O$, as illustrated in Figure 5. The April data indicate a slope of 7.486, with an R^2 value of 0.88. However, the slope of the best fit line progressively decreases to 6.19 and 6.17 in June and August, respectively, with the variance significantly increasing over this period. The initial April slope of 7.486 is identical to that of a regional meteoric water line calculated from available GNIP data and illustrated in Figure 6, whilst the June and August snowpack lines suggest a reduced slope relative to the regional line.

During snow sampling, the hardness, relative moisture content and snow grain size were assessed and logged for each snow pit profile. An example profile is illustrated in Figure 7 and indicates both the presence of icy layers, where hardness increases and relative moisture content decreases, as well as a weak correspondence between grain size and enriched $\delta^{18}O$ values. The average values for each of these physical characteristics by sampling location and month are summarised in Table 2. In general, the average hardness, moisture content and grain size all increase over time, with the largest changes occurring between April and June. These trends are consistent with processes of snowpack homogenisation and partial melting which would have taken place between the pre-melt and peak melt sampling times. All three physical characteristics, though, do also show some variation with snowpack altitude. During melt periods (June and August), snowpack hardness tends to decrease with altitude,

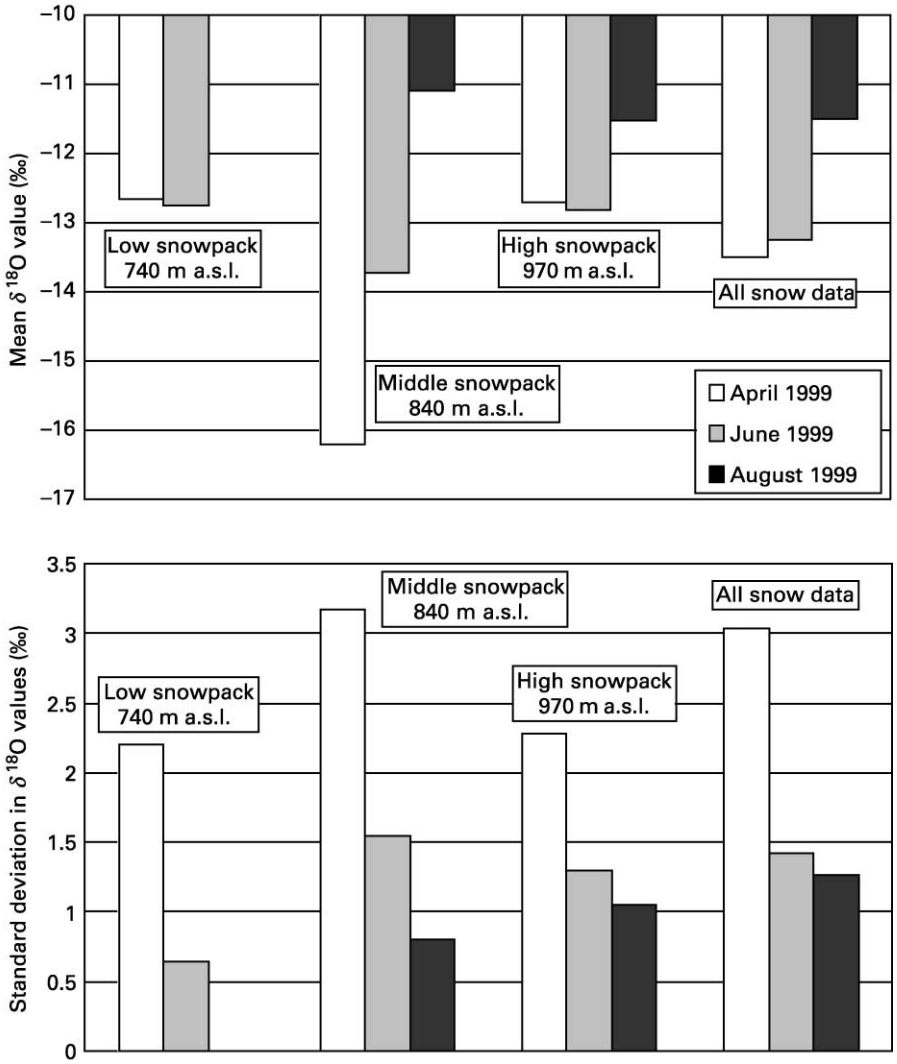

Figure 3 Mean and standard deviation in $\delta^{18}O$ values (‰) for snow samples, grouped by month and snowpack elevation

as would be expected since pack hardness can reflect processes of grain growth and ice formation during melting and these will be more advanced at the lower altitude sites. In April, however, the Middle site exhibits the most pronounced degree of snowpack hardness. This may reflect the relative location of this site, which is more exposed to high winds than the other two, resulting in the occurrence of hardened icy layers derived from surface crusting throughout the snow profile. The moisture content of the snowpacks during April decreases significantly with altitude, as does the grain size, both suggesting a more advanced snowpack evolution due to the warmer temperatures at the lower altitude. During June, altitudinal differences in moisture content and grain size are less pronounced, and by August the snowpacks at the Middle and High sites have very similar characteristics (the Low snowpack having been completely eliminated prior to August). As illustrated in Table 3, the snow profile hardness, moisture content and grain size all show weak, but statistically significant, correlations with the $\delta^{18}O$ value along the profile, while the only significant

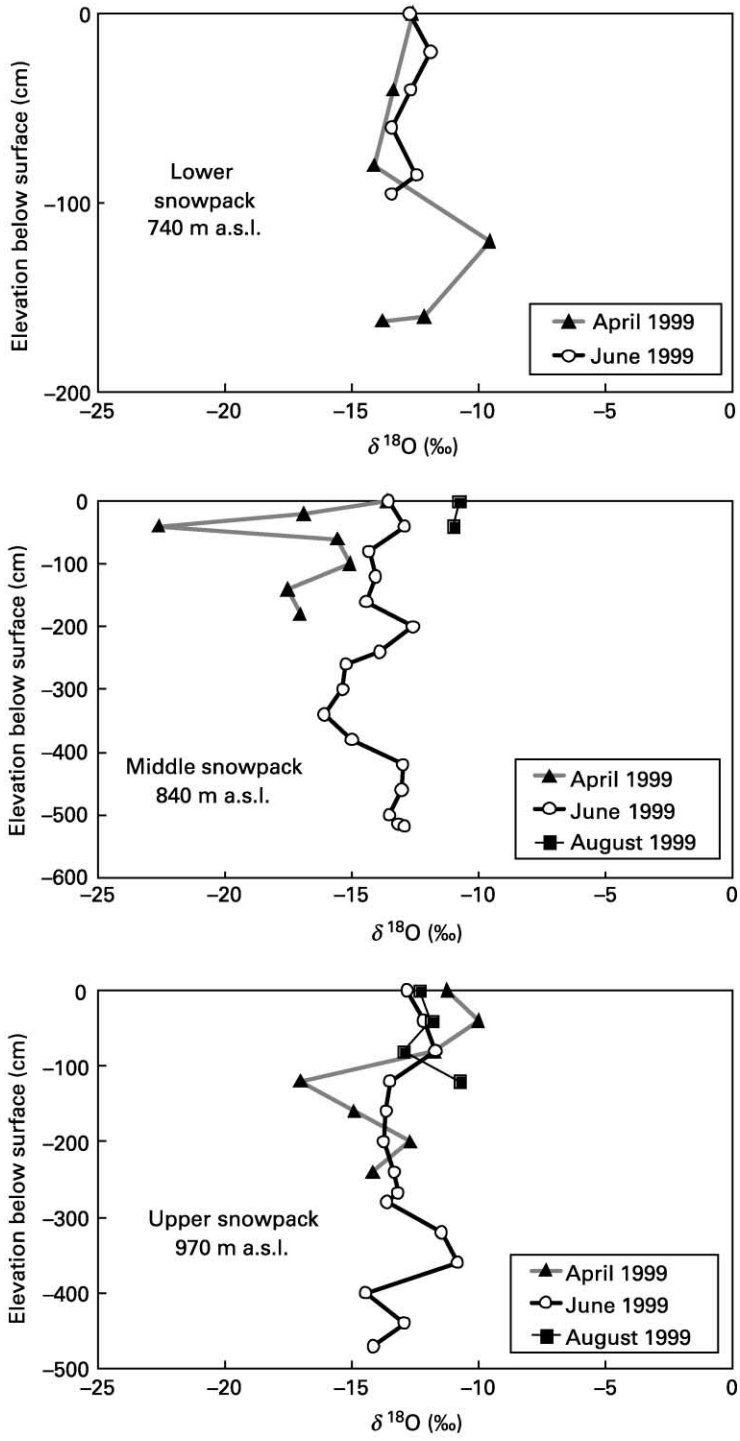

Figure 4 Snowpack isotopic stratigraphy based on $\delta^{18}O$ (‰) as a function of elevation below the snowpack surface. April profiles for the Middle and High sites are incomplete, as full excavation of the snowpack was not undertaken

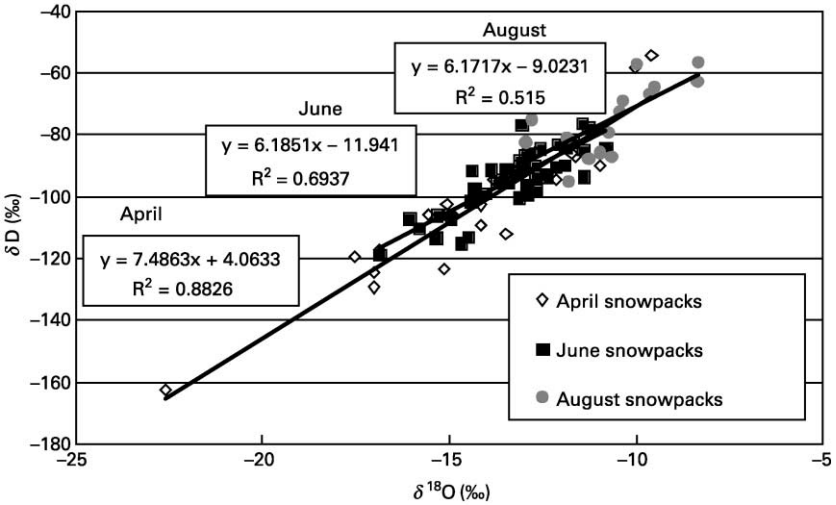

Figure 5 Plot of δD and δ^{18}O values (‰) for snowpack samples grouped by month

correlation amongst the snow physical characteristics is a strong one between moisture content and grain size.

Isotopic evolution of snow meltwater during generation and runoff

The mean values of δ^{18}O given in Table 1 suggest that the seepage at the base of the snowpack and the downstream meltwater runoff tend to be depleted in δ^{18}O relative to the source snowpack. This is most pronounced in June when the seepage and runoff have average δ^{18}O values of -14.47 and -14.28 and are derived for a snowpack source with an average δ^{18}O value of -13.26. By August, the average values are very similar, with the snowpack and seepage varying by only 0.48‰ and the downstream runoff appearing to be slightly enriched (by 0.31‰) relative to the seepage. However, these mean values represent averages of sample groups having a high degree of variability. Accordingly, difference of means tests were performed on the δ^{18}O values for samples grouped according to type and sampling month and these are summarised in Table 4. The results indicate that, in June, the average δ^{18}O value of the snowpack differs from that of the meltwater runoff, in that they

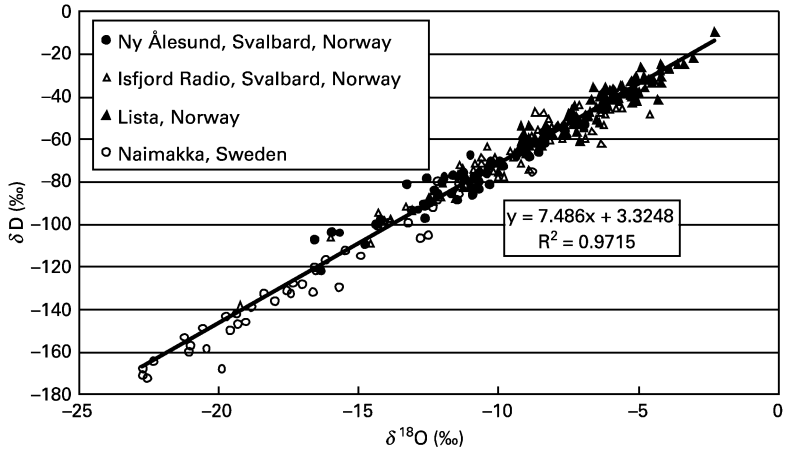

Figure 6 Regional Meteoric Water Line, estimated from GNIP data from four Scandinavian stations

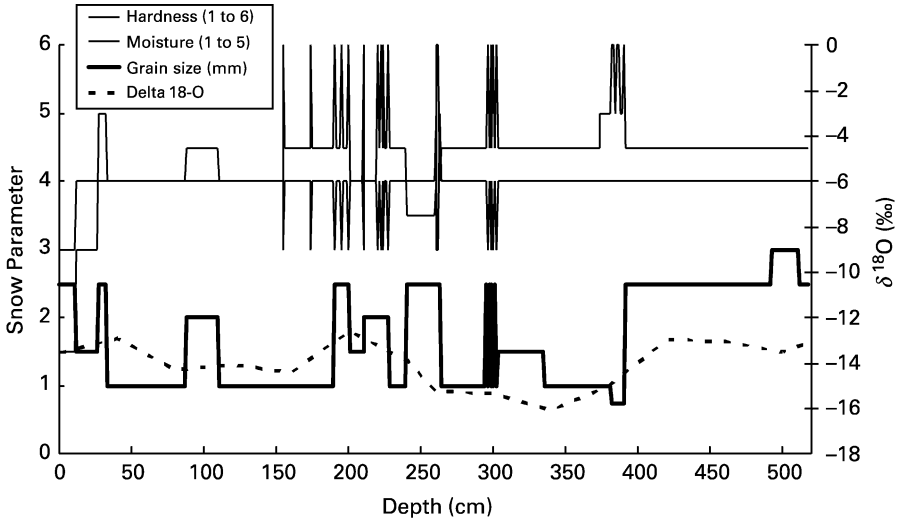

Figure 7 Vertical snow profile from the Middle site in June, illustrating the physical properties and $\delta^{18}O$ values (‰) as a function of depth below the snowpack surface

can statistically be distinguished at the $\alpha = 0.01$ significance level. The solid snowpack and snowpack seepage samples, however, cannot be distinguished. The failure to reject the null hypothesis for the difference of means between the snowpack and the seepage may reflect, in part, the higher variance and smaller sample numbers for the seepage relative to the meltwater runoff. In August, neither the snowpack seepage nor the downstream meltwater runoff can be statistically distinguished from the source snowpack. Snowpack enrichment through the progress of the snowmelt is statistically confirmed by the difference of means tests between the April and June snowpacks (significant at the $\alpha = 0.05$ level) and between the June and August snowpacks (significant at the $\alpha = 0.01$ level).

The plot of δD against $\delta^{18}O$ for the snowpack seepage and the meltwater runoff samples in June and August is illustrated in Figure 8, with the average June and August snow lines established in Figure 5 also shown, as is the best fit line for the combined seepage and meltwater data. Although the data are sparse, particularly for snowpack seepage, some interesting trends are indicated. Firstly, in virtually all cases, the isotopic compositions of the seepage samples lie below the average snow lines, suggesting a contribution from or

Table 2 Average values for snow characteristics by site and month of sampling

Month and site	Hardness (1–6)	Moisture content (1–5)	Grain size (mm)	$\delta^{18}O$ (‰)
April – Lower	3.50	3.88	1.07	−12.77
April – Middle	4.29	2.00	0.79	−17.11
April – Higher	3.57	1.00	0.31	−13.17
April average	**3.78**	**2.36**	**0.74**	**−14.27**
June – Lower	4.67	4.33	1.35	−12.64
June – Middle	4.20	3.93	1.86	−13.79
June – Higher	3.98	4.05	2.00	−13.08
June average	**4.16**	**4.02**	**1.86**	**−13.39**
August – Middle	4.33	4.00	2.33	−11.95
August – Higher	4.00	4.00	2.25	−11.85
August average	**4.17**	**4.00**	**2.28**	**−11.65**

Table 3 Correlation between physical properties estimated from snow profiles and $\delta^{18}O$ content of snow samples. The first number in each box indicates the correlation coefficient and the second number is the t statistic (based on $n = 86$). Significant correlations at the 0.05 level (critical $t = 1.662$) are highlighted in bold

	Moisture content	Grain size	$\delta^{18}O$
Hardness	0.063	0.001	**0.220**
	0.58	0.005	**2.07**
Moisture content		**0.631**	**0.229**
		7.49	**2.16**
Grain size			**0.206**
			1.95

combination with a source that has been subjected to evaporation. This is especially of interest in the case of snowpack seepage, as significant interaction with soil seepage water is unlikely at that point along the flow path. The meltwater runoff data indicate possible cases of both deuterium depletion and its excess, although the magnitude of depletion, where present, is much greater. The paucity of data and high variance, though, do prevent one from making definitive conclusions from the individual data sets for seepage and meltwater. However, the best fit line for the combined seepage and meltwater runoff values has a slope of 4.4 and is significant at the 0.05 level, indicating a decrease in slope from the June "snow line". Figure 9 also illustrates the isotopic shift between the snowpack and the seepage for average values at individual sampling sites during June and August 1999, as well as for August 1998. These data, though again based on a limited number of samples, illustrate the expected depletion that occurs when meltwater is generated from a snowpack. They also suggest that, in June, melting is associated not only with isotopic depletion, but also produces a meltwater with relatively less δD, such as is associated with waters subjected to evaporation.

Short-term isotopic variability along a meltwater flow path

The diurnal pattern of isotopic evolution along a flow path was briefly considered at the Middle site during early June 1999. During and following the excavation of snow sampling pits at this site, successive downstream points along the path of meltwater were sampled every four hours over 28 hours, and all flow points were also sampled at 1200 hours over a six-day period. The sampling points along the flow path are illustrated in Figure 2 and comprise (1) snow pit in snowpack; (2) snowpack seepage water at the edge of the snowpack, 60 m down gradient from the snow pits; (3) snow meltwater on soil surface, 3 m below the edge of the snowpack; (4) soil seepage water sampled 15 cm below ground surface and 5 m

Table 4 Results of the difference of means tests for June and August snowpacks and runoff. The numbers indicate the mean difference and those in bold were found to be statistically significant at the $\alpha = 0.01$ level

	Snowpack June 1999	Seepage August 1999	Runoff June 1999	Runoff August 1999
Snowpack August 1999	**1.79**	0.479		0.207
Seepage June 1999	0.945	2.24	0.227	
Seepage August 1999				0.686
Runoff June 1999	**1.17**			**3.15**

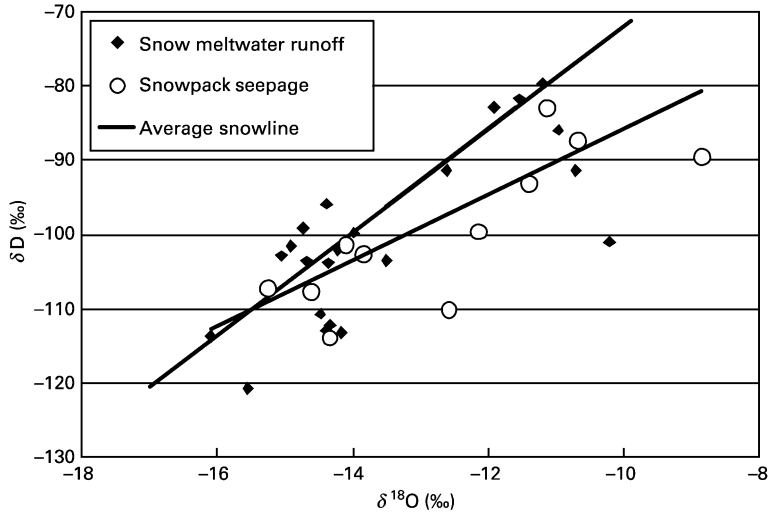

Figure 8 Plot of δD and δ^{18}O values (‰) for snow meltwater runoff and snowpack seepage. The best fit June "snow line" from Figure 5 ($y = 6.19x - 11.94$) is also shown

downstream of the previous sampling point; (5) channelled meltwater runoff approximately 35 m below sampling point 4; and (6) channelled meltwater runoff at the point of discharge into Okstindtjønna, approximately 100 m below sampling point 5. The travel time of snow meltwater along this pathway is believed to be relatively short (< 60 min), especially during periods of relatively rapid melt (for example, when air temperatures are highest in the middle of the day).

The variation in δ^{18}O along this flow path is illustrated in Figure 10 and suggests a relatively stable pattern between the snowpack and the depleted snowpack seepage and downstream meltwater runoff. The soil surface and subsurface samples, however, do show a higher degree of variability, though, with some particularly depleted values found on the soil surface at 0000 and 0800 hours on 9 June 1999. Given the stable pattern in the isotopic composition of the runoff sampled downstream from these points, though, the soil seepage water contributions are apparently quite localised. In fact, the average meltwater runoff values are slightly enriched relative to the snowpack seepage and this cannot be explained by

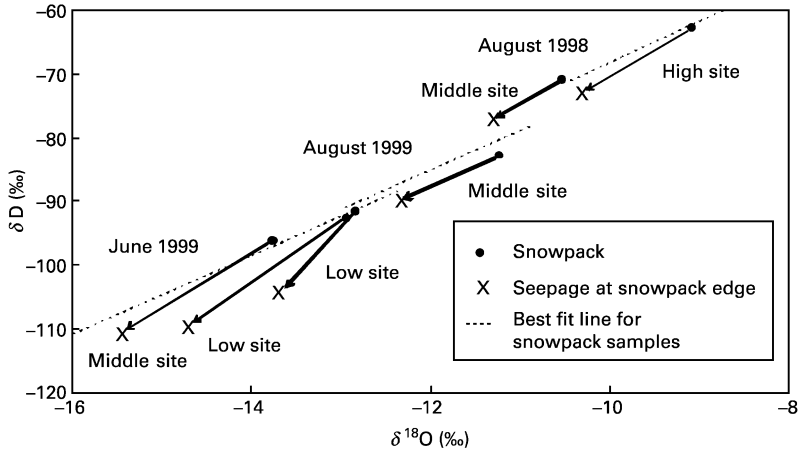

Figure 9 Observed isotopic shift in δD and δ^{18}O values (‰), comparing average snowpack values for a given site with those of the seepage generated by the snowpack

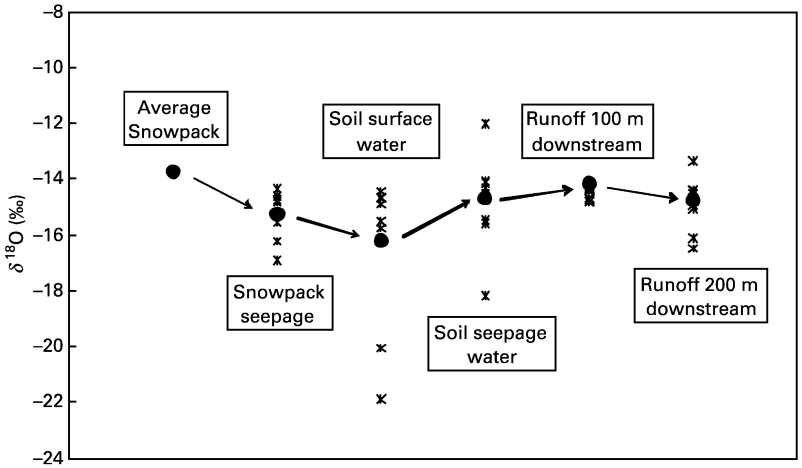

Figure 10 $\delta^{18}O$ values (‰) along a flow path at the Middle site during June 1999 intensive sampling over a five-day period. The solid circles indicate average values

interaction with the depleted soil moisture, but may reflect further evaporation of the runoff following seepage or contributions for rainfall (Table 1). Figure 11 illustrates the variability between sampling sites along the flow paths as a function of time and suggests two trends. Firstly, the snowpack seepage is most depleted at 1200 and 3600 hours, which may reflect a diurnal peak in meltwater production. Secondly, at 2400 hours, the highest degree of variability along the flow path is observed and this may correspond to the diurnal low in seepage discharge, such that the contribution of earlier, more depleted, soil seepage water makes a more significant local contribution along the flow path.

Discussion and conclusions

The mean values for 1999 snowpack at Okstindan (Table 1) are similar to, but slightly more depleted than, the multi-year average values reported by Theakstone and Knudsen (1996a)

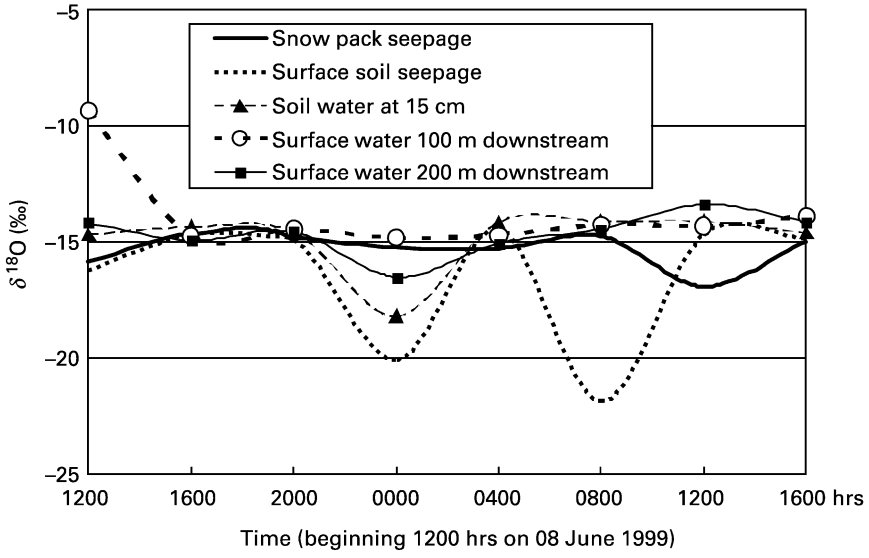

Figure 11 Diurnal variations in $\delta^{18}O$ values (‰) at each sampling site along a flowpath downstream of the Middle snowpack. Samples were taken every four hours over a 28 hour period

for an adjacent catchment. The overall trends in the stable isotopic composition of the snowpack through the melt season seen in our data are also similar to those reported in other recent studies. In particular, the decrease in the variability of $\delta^{18}O$ values associated with snowpack isotopic has also been reported by Taylor *et al.* (2001) and Unnikrishna *et al.* (2002) for Sierra Nevada snowpacks and by Theakstone and Knudsen (1996b) in the Okstindan region. Changes in the isotopic stratigraphy associated with snowpack homogenisation have also been previously observed (e.g. Sturm and Benson 1997; Taylor *et al.* 2001), although earlier studies have generally been undertaken within well-controlled field experimental conditions. As shown here, this snowpack homogenisation is also associated with an overall enrichment in the $\delta^{18}O$ values, although the level of enrichment observed in this study was much greater between the June and August sampling periods, whilst the decrease in variability in $\delta^{18}O$ was much more pronounced between April and June. This suggests that snowpack processes, such as melt percolation, taking place between the pre-melt and peak melt sampling stages served to largely homogenise the isotopic stratigraphy, whilst the subsequent melting of this much more homogeneous snow then progressively contributed to the enrichment of the residual snow. Processes of snowpack evolution and meltwater percolation are also seen in the evolution of the physical properties of the snowpack in the form of a progressive increase in hardness, moisture content and average grain size through the melt season. Significant correlations between physical properties and isotopic values for individual snow pit samples, however, are very weak. This may be a consequence of the relatively qualitative and subjective methods used to acquire data on physical properties in the field, but would also reflect the fact that much of the evolution of isotopic values at a point in the profile is dependent upon the past history of meltwater and vapour transfer to that point. In contrast, changes in physical properties would depend more directly on snowpack temperature at that point.

The $\delta^{18}O$ values of meltwater increased to an even greater extent than did the values for the snowpack between the peak melt and residual melt sampling periods. This increase in isotopic values in meltwater through the melt season has also been previously reported (e.g. Hooper and Shoemaker 1986; Cooper *et al.* 1993; Taylor *et al.* 2001; Unnikrishna *et al.* 2002) and clearly presents a problem when using a constant isotopic value for the snowmelt contribution in isotopic hydrograph separations. The results presented for our sites also indicate that, during the June peak melt season, the snowpack isotopic signature can be statistically distinguished from the snow meltwater it generates, reflecting the preferential removal of lighter water from the snowpack. By August, however, the meltwater signature can no longer be differentiated from that of the snowpack, pointing to the more complete melting of a homogeneous snowpack core late in the season.

In this study, a subset of the samples were analysed for values of δD. A number of workers have adopted a similar approach, for example, Hooper and Shoemaker (1986) and Cooper *et al.* (1993), although in general, previous work on snowpack and meltwater isotopes has tended to focus largely on the use of $\delta^{18}O$. Hooper and Shoemaker (1986) analysed δD only to evaluate whether evaporation has taken place during meltwater sampling, whereas Cooper *et al.* (1993) recognise that trends in the plot of δD against $\delta^{18}O$ may highlight the effects of evaporation or sublimation on the snowpack. The analyses of δD together with $\delta^{18}O$ allow two interesting observations to be made, which have received little previous attention. Firstly, by compiling GNIP data to establish an RMWL (Figure 6) and comparing this with the "snow lines" for each sampling period, trends in the slope of the line as the melt season progresses can be determined. Our data indicate that the initial (April) slope of the snow line is identical to that of the RMWL, but that the snowpacks during June and August are characterised by a decrease in the slope of this line. The largest drop in the slope occurs between April and June, when the slope

changes from its initial value of 7.49 to a June value of 6.18. This change in slope implies that the processes of melt metamorphism and snowpack changes over time involve not only an overall homogenisation, but also the preferential removal of hydrogen over oxygen isotopes, as is associated with evaporation and sublimation. Secondly, the snowpack seepage and meltwater samples also show a significant drop in the slope of the δ^{18}O-δD line when compared with the snowpack from which they were generated. Although it might be anticipated that snow meltwater runoff would be subjected to some evaporation, due to the very shallow character and low velocities of flows that seep from the snowpack at its edge, the fact that seepage sampled from the snowpack prior to discharge also exhibits this drop in slope is intriguing. Further work is clearly warranted, but these results suggest that evaporative processes, both at the top of the snowpack or via vapour transport within the snowpack, may be at work (see Albert and McGilvary 1992; Friedman *et al.* 1991; Sommerfield *et al.* 1991). When deuterium excess values are calculated for the snowpacks (that is, subtracting the δD predicted by the RMWL from the observed δD, as shown in Figure 12) it is clear that evaporative processes in the upper portion of the snowpack are occurring. It may be that the seepage samples represent the preferential melting of these surface layers and are thus relatively depleted in δD. Additionally, once meltwater is generated within the snowpack, it may be subjected to further local evaporation during migration within the snowpack, due to vapour transport and convection within the pack (Sturm and Johnson 1991). An alternative explanation would be that mixing with soil seepage water previously subjected to evaporation contributes to this signature, thus modifying the isotopic values even before they reach the edge of the pack. However, as the snowpacks on the Middle and High sites overlie a minimal soil cover (their late-lying character inhibiting soil development), this seems less likely.

Analyses of the short-term pattern of isotopic variability along a flow path (Figures 10 and 11) suggest that the widest range of isotopic values along the flow path occurs during the night, when meltwater generation and surface flow rates are assumed to be much lower. Taylor *et al.* (2001) considered the relationship between flow rate and δ^{18}O values within a snowpack during meltwater generation and found that low flow rates produce lower δ^{18}O values, thus corroborating previous theoretical work by Árnason (1969) and experimental work by Herrmann *et al.* (1981). Although we are considering a much larger and more complex system here, Figure 10 does indicate that the lowest values of δ^{18}O for

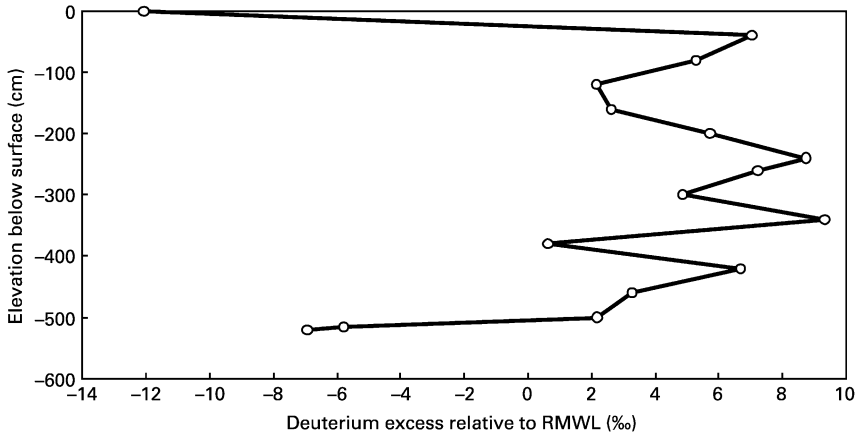

Figure 12 Deuterium excess in snow profile at the Middle site sampled during June. The excess is calculated from δD (observed) - δD (predicted from RMWL)

the runoff 200 m downslope of the snowpack occurred at 2400 hours, whereas the highest values are associated with 1200 hours. However, the soil seepage water exhibits the most complex pattern of variability along the flow path and often has isotopic values much lower than the meltwater. The temporal pattern implies a variable degree of mixing between the snow meltwater and the soil seepage water through a diurnal cycle, with the lower isotopic values possibly reflecting the contribution of residual water from early snowmelt (which would have been more depleted) and the more normal values occurring when meltwater generation and runoff rates are high and dominate flow through the soil seepage water sampling sites.

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