δD–δ^{18}O relationships on a polythermal valley glacier: Midtre Lovénbreen, Svalbard

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This paper outlines the results of stable isotope (δD–δ^{18}O) analysis of snow and glacier ice undertaken as part of a larger study concerning structural glaciology, debris entrainment and debris transport patterns at Midtre Lovénbreen, Svalbard. Samples of fresh snow were collected from the glacier surface in spring 1999 and samples of surface glacier ice and basal ice samples were collected in summer 1999. When plotted on bivariate co-isotopic diagrams (δD–δ^{18}O), the slopes obtained for snow and unmodified glacier ice (6.4 and 6.9, respectively) are less steep than those for the basal ice layer and transverse ice layers on the ice surface (7.6 and 7.7, respectively). The difference in the slope of these lines is not statistically significant at the sample size (50) used in this study. The results indicate that although stable isotope analysis clearly has potential for studies of debris entrainment, transport and structural glaciology, difficulties remain with applying this technique. It is therefore not possible to apply these isotopic techniques to ice facies of unknown origins. In particular, large sample numbers are required to establish statistical differences and high-resolution sampling of specific ice facies may be necessary to establish isotopic differences.

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Theoretical and empirical studies indicate that δD–δ^{18}O relationships represent a powerful analytical tool in glaciology, in particular in the interpretation of the thermal history of glacier ice and snow, rock glacier ice, and in studies of debris entrainment by glaciers (MacPherson & Krouse 1967; Ambach et al. 1968; Jouzel & Souchez 1982; Souchez & Jouzel 1984; Souchez & de Groote 1985; Sugden et al. 1987a, 1987b; Gordon et al. 1988; Knight 1989, 1994; Raben & Theakstone 1994; Hubbard & Sharp 1995; Lawson et al. 1998; Humlum 1999). Since water undergoes isotopic fractionation on freezing, combined δD and δ^{18}O measurements make it possible to distinguish between isotopically-unmodified surface glacier ice and isotopically-modified basal ice by plotting values of δD and δ^{18}O deuterium and oxygen in ‰ relative to SMOW) on a co-isotope diagram (Souchez & Jouzel 1984; Lehmann & Siegenthaler 1991).

δD–δ^{18}O relationships have been used previously to interpret the thermal history of glacier ice and to determine whether basal melting and refreezing has taken place in a given sample of ice (Arnason 1969; Jouzel & Souchez 1982; Souchez & Jouzel 1984; Souchez & de Groote 1985; Gordon et al. 1988; Knight et al. 1994). Samples of glacier ice that have not undergone refreezing lie on the same regression line as precipitation (normally with a slope of around 8) when plotted on a δD–δ^{18}O diagram. Ice that has undergone refreezing produces a regression line with a lower

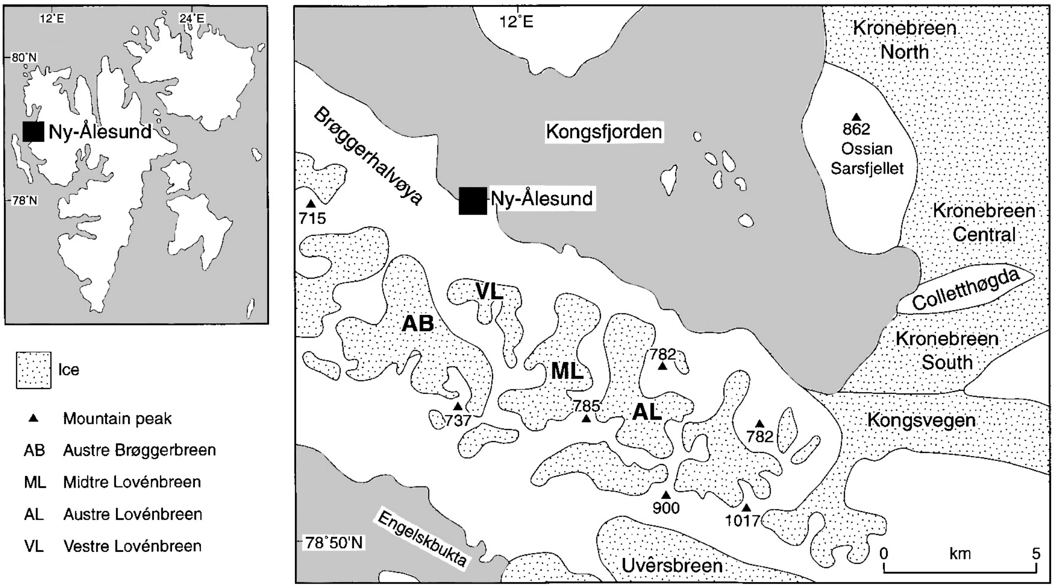

Fig. 1. Location map of the peninsula of Brøggerhalvøya, north-west Spitsbergen. The glacier studied, Midtre Lovénbreen, and the settlement of Ny-Ålesund are indicated.

slope, depending on the initial δ values of the melted ice at the onset of refreezing.

δD–δ^{18}O relationships have clear but as yet untested applications to structural glaciology. Studies of this nature have only been undertaken on temperate glaciers (Epstein & Sharp 1959; Sharp et al. 1960; Hambrey 1974; Lawson & Kulla 1978). The technique has never been applied to Svalbard glaciers, where stable isotope analysis has the potential to distinguish between those areas of a glacier where basal melting has occurred and those where it has not. This allows us to test theories concerning the origin of different glacigenic structures including the debris-rich transverse surface fractures present near the termini of many Svalbard glaciers. On the basis of detailed surface and three-dimensional structural mapping, and their association with basal debris, these transverse surface fractures have previously been interpreted as thrusts of basal origin (Hambrey et al. 1996, 1997, 1999; Glasser et al. 1998; Glasser & Hambrey 2001). However, the extent to which basal ice is involved in thrusting is unknown. The work reported here is part of a wider programme of research concerning the structural glaciology, dynamics, debris transport processes and landform development at high Arctic valley glaciers.

Study area and sampling methods

Study area

The Svalbard archipelago lies between 77° and 80° N. The archipelago is currently 60 % glacierized and is dominated by a maritime Arctic climate (Hagen et al. 1993). The glacier examined in this paper, Midtre Lovénbreen, is a small (4.2 km long, 5.95 km^2 in area) valley glacier on Brøggerhalvøya (Brøgger Peninsula), near the settlement of Ny-Ålesund in north-west Spitsbergen (Fig. 1). Midtre Lovénbreen is fed by accumulation areas in four tributary cirques (Fig. 2) and is largely crevasse-free except in areas where basal topography changes abruptly and where tributary glaciers enter the main flow unit from the steep upper basins. Measured velocities on the centre-line of Midtre Lovénbreen at the equilibrium line range from 4.4 ma^{-1} (Bjornsson et al. 1996) to 7.3 ma^{-1} (Liestøl 1988).

Norwegian Polar Institute measurements on Midtre Lovénbreen show a continuous negative mass balance since 1967 (Hagen & Liestøl 1990). Statistical analysis of mass balance records and climate data suggests that the net mass balance has probably been negative in the majority of years since 1900 (Lefauconnier &

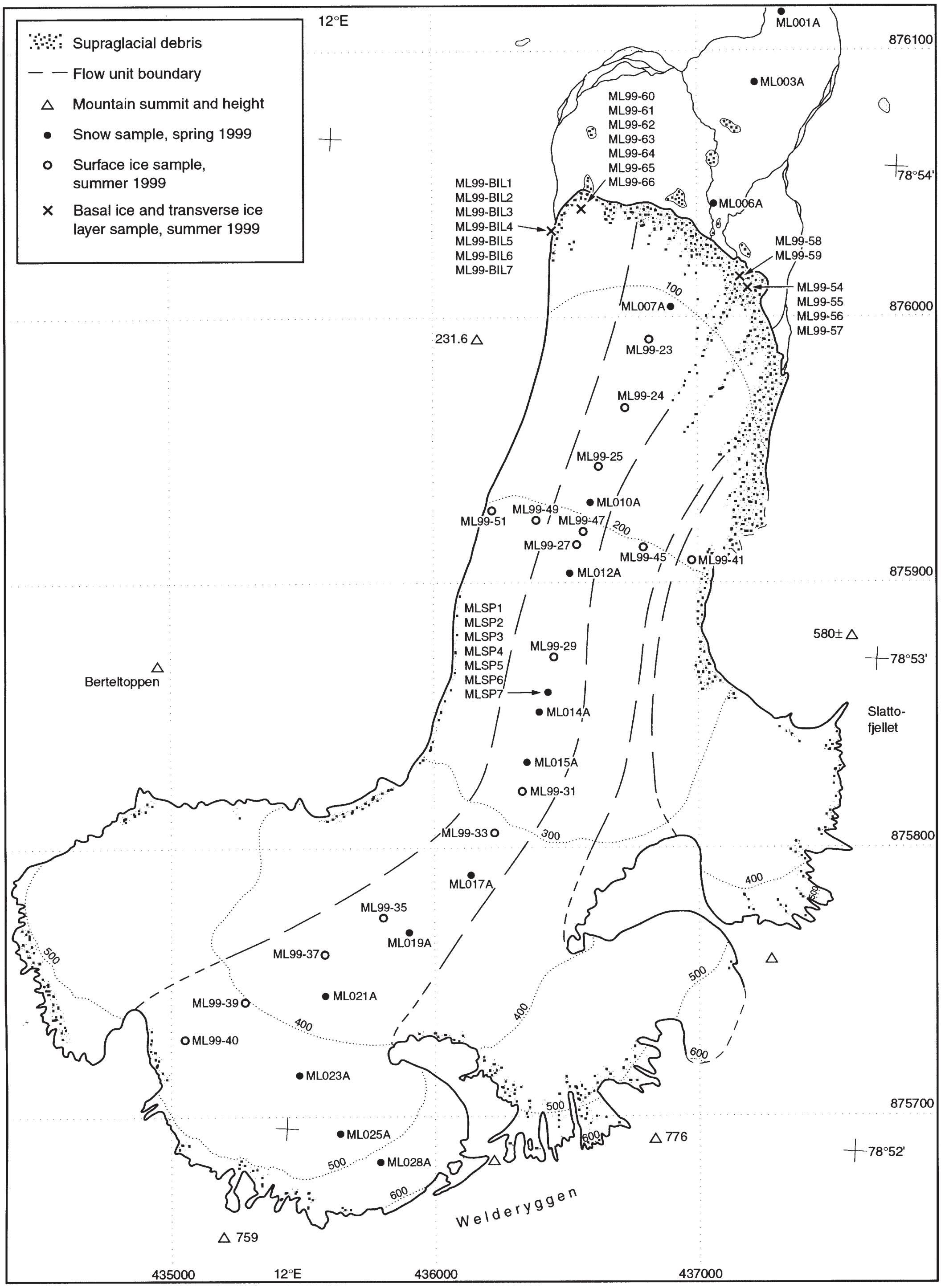

Fig. 2. Map of Midtre Lovénbreen showing the location of samples collected in spring 1999 (snow samples) and summer 1999 (surface glacier ice and basal ice/transverse ice layer samples).

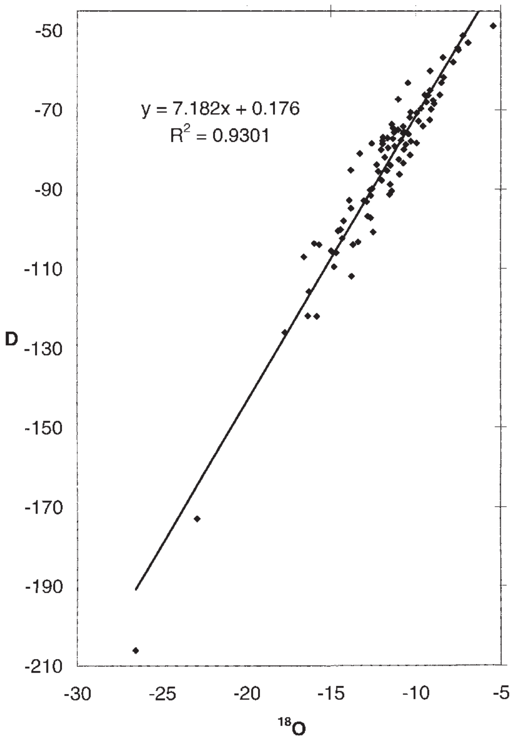

Fig. 3. Bivariate co-isotopic plot (δD–δ^{18}O) of monthly precipitation recorded at Ny-Ålesund from January 1990 to December 1997. The data source for this diagram is the IAEA Global Network of Isotopes in Precipitation (GNIP) database (available on the Internet at http://www.iaea.org/programs/ri/gnip/gnipmain.htm).

Hagen 1990; Lefauconnier et al. 1999). Glaciers on Brøggerhalvøya are currently receding from their Neoglacial maxima (ca. 1890), and Midtre Lovénbreen has receded almost 1 km since this time (Hagen et al. 1993). Volume losses since the post-glacial maximal position have been substantial, possibly as much as 25%, on the basis of former ice marginal positions and trim-lines (Hamberg 1894; Liestøl 1988; Hansen 1999). Radio-echo soundings of the internal temperature characteristics and bed topography of Midtre Lovénbreen indicate that the glacier is polythermal, with extensive areas of temperate ice beneath the accumulation areas but with a terminus that is frozen to the bed (Hagen & Saetrang 1991; Hagen et al. 1991; Ødegård et al. 1992; Björnsson et al. 1996).

The International Atomic Energy Agency (IAEA) provides details of isotopic contents (including oxygen-18, deuterium and tritium) of composite monthly precipitation on their Global Network of Isotopes in Precipitation (GNIP) database. Monthly values from this database (January 1990 to December 1997) from Ny-Ålesund (location 78° 15' N, 11° 56' E; altitude 7 m a.s.l.) plotted on a δD–δ^{18}O diagram (Fig. 3) show a linear relationship defined by the least-squares regression line δD = 7.182δ^{18}O + 0.176 (r^2 = 0.93). Although these data hide considerable variation in seasonal precipitation, they indicate a local meteoric line at Ny-Ålesund with a slope of 7.2 (Fig. 3). Unfortunately, since these data are collected at sea level, they provide no information about the altitudinal variation of the isotopic composition of snowfall on the glacier.

Sample collection and analysis

To characterize changes in the isotopic composition of snow with altitude, samples were collected from Midtre Lovénbreen over two consecutive days in spring 1999 (29 and 30 April 1999). Temperatures during this period ranged from -16 °C to -6 °C and there was neither snowmelt nor significant snowfall during the period of sample collection. Sample collection followed the centre-line of the glacier, rising from sea level (for spring snow) and the glacier snout (for ice samples) to an altitude of 534 m a.s.l. in the accumulation area of Midtre Lovénbreen (Fig. 2). Altitudes at each sample point on the glacier were determined using a hand-held Thommen altimeter calibrated daily at sea level. Revisits to sea level at the end of sample collection on both days provide an estimate of the accuracy of the altimeter of ±4 m. At each sample site, fresh surface snow was removed from the glacier surface and samples were collected from an older, firmer snowfall at 5-20 cm depth in the snowpack. Snow samples were allowed to melt completely in airtight bottles in the shade before being transferred to clean airtight 8 ml glass vials to prevent evaporation. Summer sample collection was performed in August 1999 when the glacier surface was largely snow-free and bare glacier ice was exposed to an altitude of 440 m a.s.l. in the upper basin of Midtre Lovénbreen. Samples of surface glacier ice were collected along the centre-line from sea level to the upper basin. Specific ice facies, including glacier ice in transverse surface fractures previously interpreted as thrusts and basal ice exposed in a lateral cliff section, were also sampled. Ice facies descriptions follow those of Knight (1989, 1994).

Oxygen and hydrogen isotope ratios were measured according to standard procedures at the NERC Isotope Geosciences Laboratory in Keyworth, UK (Heaton 1992; Arrowsmith 1998). D/H ratios were determined on H_2 produced from 4 µl samples of water reacted in vacuo with ca. 120 mg of zinc turnings for 40 min at 500 °C. $^{18}O/^{16}O$ ratios were determined on CO_2 equilibrated with 2 ml samples of water at 23 °C for 6 h in a VG Isoprep 18. The D/H and $^{18}O/^{16}O$ ratios were measured in a VG SIRA II mass spectrometer and compared with the ratios for two laboratory standards analysed at the same time. The standards are calibrated against VSMOW and SLAP, and sample ratios are reported in ‰ as δD and $\delta^{18}O$ values versus VSMOW. Analytical precision is better than ±2 ‰ for δD and ±0.1 ‰ for $\delta^{18}O$. All regression calculations were performed in Excel using the Pearson product moment correlation.

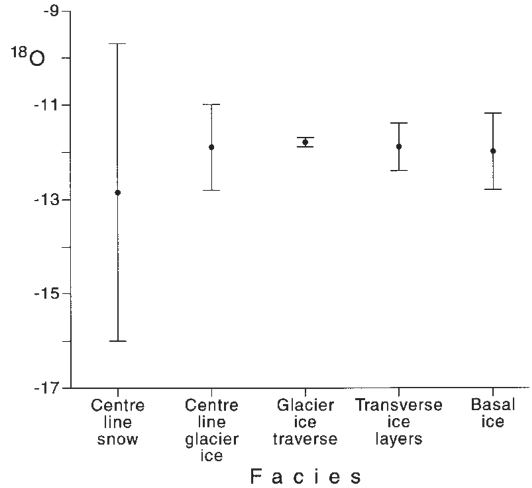

Fig. 4. Mean and standard deviation of $\delta^{18}O$ in the snow and ice facies identified at Midtre Lovénbreen.

Results: δD–$\delta^{18}O$ relationships

δD values for surface snow on Midtre Lovénbreen range between -54 and -126 ‰, with a mean value of -88.8 ‰. $\delta^{18}O$ values range between -6.8 and -18.6 ‰, with a mean value of -12.81 ‰ (Fig. 4). When plotted on a δD–$\delta^{18}O$ diagram (Fig. 5a) the surface snow measurements show a linear relationship defined by the least-squares regression line $\delta D = 6.42\delta^{18}O - 6.49$ ($r^2 = 0.96$). δD values for surface glacier ice on Midtre Lovénbreen range between -70 and -90 ‰, with a mean value of -84.2 ‰. $\delta^{18}O$ values range between -10.1 and -12.8 ‰, with a mean value of -11.94‰ (Fig. 4). When plotted on a δD–$\delta^{18}O$ diagram (Fig. 5b) the surface glacier ice measurements show a linear relationship defined by the least-squares regression line $\delta D = 6.92\delta^{18}O - 1.47$ ($r^2 = 0.96$). Although the means of both the snow and glacier ice data sets are similar, the glacier ice data set shows a greater clustering of sample points around the mean than the snow data set (Fig. 5a, b).

Variations in δD and $\delta^{18}O$ within the snowpack are apparent with altitude (Fig. 6), although there is no clear trend in this variation and nothing to indicate that the snow becomes isotopically lighter with altitude. A snowpit was dug at an altitude of ca. 260 m a.s.l. on Midtre Lovénbreen to investigate variations in $\delta^{18}O$ and δD with depth in the snowpack (Table 1). These data indicate that the values of $\delta^{18}O$ and δD vary considerably with depth in the snowpack and that the individual snow facies have different isotopic signatures.

A lateral melt stream flowing along the north-west margin of Midtre Lovénbreen exposes a ca. 5 m high section of basal ice resting on a layer of frozen sediment. Here low-angle thrusts

Table 1. Snow stratigraphy and associated δD–$\delta^{18}O$ values in a snowpit at 260 m on Midtre Lovénbreen, spring 1999.

Depth (cm)	Snow characteristics	Interpretation	$\delta^{18}O$ v SMOW	δD v SMOW
0 - 8	Soft snow	Fresh snowfall	-9.56	-66.5
9 - 32	Firm snow	Recent snowfall	-10.1	-66.5
33 - 72	Coarse, crystalline snow	Compacted snow with depth hoar	-15.24	-117.6
73 - 74	Thin ice layer	Depth hoar?	-7.49	-52.2
75 - 90	Sugary snow with ice layers	Depth hoar?	-10.19	-65.7
91 - 100	Clear ice layer	Superimposed ice	-14.18	-104
101 - 104	Bubbly ice layer	Glacier ice	-11.11	-83.6

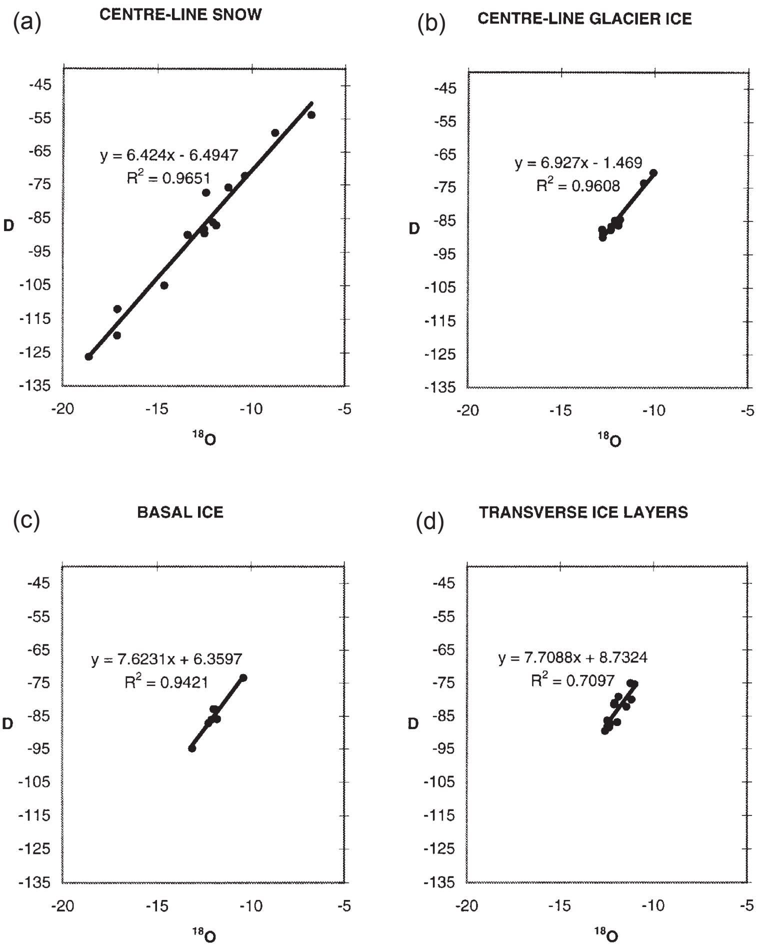

Fig. 5. Bivariate co-isotopic plot (δD–δ¹⁸O) of samples from Midtre Lovénbreen: (a) fresh snow sampled along the centre line of the glacier; (b) surface glacier ice sampled along the centre line of the glacier; (c) basal ice sampled in lateral ice cliff; (d) transverse ice layers on the glacier surface.

penetrate up into the basal ice layer from the sediment beneath. The thrusts steepen upwards to break the glacier surface at angles of 30 - 40°, where they are commonly associated with piles of debris on the glacier surface. Thrusts typically comprise alternating bands of coarse clear and coarse bubbly ice. The coarse clear ice facies are often associated with disseminated sandy gravel layers and isolated pebble-sized clasts. Interstitial mud films and disseminated mud clots are also common within the coarse clear ice facies. δD values for basal ice exposed in this lateral cliff section range from -73 to -95 ‰, with a mean value of -84.8 ‰. $\delta^{18}O$ values range between -10.4 and -13.1 ‰, with a mean value of -11.95 ‰ (Fig. 4). When plotted on a δD–$\delta^{18}O$ diagram (Fig. 5c) the basal ice measurements show a linear relationship defined by the least-squares regression line $\delta D = 7.62\delta^{18}O + 6.35$ ($r^2 = 0.94$).

A series of transverse ice layers is present across the surface of the glacier in its ablation area. On other Svalbard valley glaciers, these features have been interpreted as thrusts (Hambrey et al. 1999). On Midtre Lovénbreen many of these transverse ice layers are debris-rich and associated with debris cones and mounds on the glacier surface. Ice facies within these transverse ice layers are locally variable but typically comprise either a layer of coarse-clear ice of variable thickness or a band of debris with interstitial glacier ice, bounded on either side by coarse-bubbly ice. In many instances the coarse-clear ice is associated with an interstitial mud film, disseminated mud clots and clotted ice facies. Samples from 12 transverse ice layers on the glacier surface show $\delta^{18}O$ values in the range -11.01 to -12.56 and δD values in the range -75.1 to -89.6 (Table 2). δD values for glacier ice in the transverse ice layers on the ice surface interpreted as thrusts range between -75 and -90‰, with a mean value of -82.8‰. $\delta^{18}O$ values range between -11.0 and -12.6‰, with a mean value of -11.87‰ (Fig. 4). When plotted on a δD–$\delta^{18}O$ diagram (Fig. 5d) the transverse ice layer measurements show a linear relationship defined by the least-squares regression line $\delta D = 7.71\delta^{18}O + 8.73$ ($r^2 = 0.71$).

Discussion

The surface glacier ice samples define a δD–$\delta^{18}O$ slope that closely approximates the local meteoric line obtained from the snow samples (6.9 and 6.4, respectively). The slope of the surface glacier ice samples (6.9) is similar to the slope of 7.2 obtained from the monthly IAEA data, suggesting that the surface ice on Midtre Lovénbreen has not undergone significant isotopic modification during the transformation from snow to glacier ice. There is some evidence of isotopic homogenization during the transformation from snow to glacier ice since the extreme values present in the snowfall are lost from the surface ice on a co-isotopic plot (Fig. 5a, b). Values of $\delta^{18}O$ and δD vary considerably with depth in the snowpack, suggesting that there is considerable variation in the isotopic composition of the snow between individual snowfall events. There may also be some isotopic modification of the snowpack during the transformation from snow to ice, perhaps through the percolation of surface meltwater or from melting/refreezing events within the snowpack itself. Since there is no clear indication that the snow becomes isotopically lighter with altitude on the glacier, it may be that (i) there is little altitudinal variation in snowfall, (ii) there is homogenization by blowing snow, (iii) the snow samples collected in spring 1999 represent multiple snowfall events. The slope of the glacier ice samples (6.9) is close to the slope of 7.2 obtained from the monthly IAEA data, which provides an average of the isotopic values in precipitation at Ny-Ålesund, giving some indication as to the isotopic content of unmodified snow and glacier ice.

Samples obtained from the basal ice layer and transverse ice layers on the ice surface plot on a line with a steeper slope than those for surface snow or unmodified glacier ice. However, the isotopic slopes of the basal ice and transverse ice layers (7.6 and 7.7) are not significantly different from those of the surface snow or surface ice (6.4 and 6.9) at the 95% confidence level. Therefore isotopic differences cannot be used to differentiate between basal ice and unmodified glacier ice on Midtre Lovénbreen. Potential reasons for this include the relatively

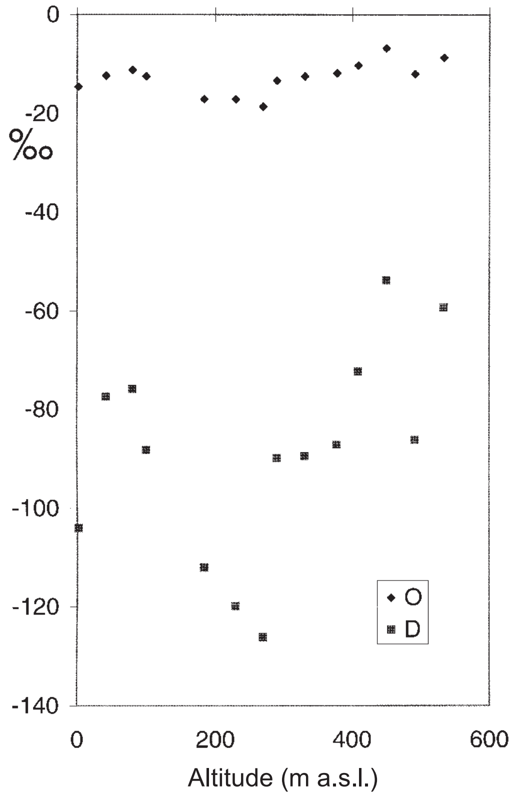

Fig. 6. Altitudinal variation in δD and $\delta^{18}O$ in surface snow on Midtre Lovénbreen.

Table 2. Ice facies in 12 transverse ice layers on the glacier surface and associated δD–$\delta^{18}O$ values, summer 1999.

Ice facies	$\delta^{18}O$ v SMOW	δD v SMOW
Coarse-bubbly ice	-11.01	-75.4
Coarse-bubbly ice	-11.16	-80.1
Coarse-bubbly ice	-12.31	-87.3
Coarse-clear ice	-11.86	-79.2
Coarse-clear ice	-11.91	-86.9
Coarse-clear ice	-12.45	-86.4
Coarse-clear ice	-12.56	-89.6
Coarse-clear ice with debris	-12.10	-81.5
Coarse-clear ice with mud clots	-11.43	-82.2
Clotted ice	-11.21	-75.1
Debris with interstitial ice	-12.06	-81.1
Debris with interstitial ice	-12.35	-88.5

small number of samples analysed, the possibility of diagenetic fractionation and uncertainties concerning the relative importance of open-system and closed-system freezing conditions. For example, the isotopic signature in precipitation may not be retained during the transformation of snow to glacier ice if there is significant surface melting and/or meltwater percolation through the snowpack. At present, we do not know the extent to which this happens, although it is commonly assumed that surface melting is limited on high Arctic glaciers in comparison to temperate glaciers. High latitude (i.e. "dry-snow") snowpacks are therefore more likely to retain the seasonal pattern of variations in $\delta^{18}O$ and δD during transformation of snow to glacier ice. Both altitudinal and across-glacier variations in $\delta^{18}O$ and δD are poorly constrained at present. There is clearly a need for wider spatial coverage of snow on the glacier surface to establish the isotopic variation in snow with altitude. Equally, it is difficult to draw firm conclusions about the variations in $\delta^{18}O$ and δD in the surface glacier ice and basal ice/transverse ice layer samples at the spatial sampling resolution used. Further research, including high-resolution sampling around individual basal ice layers and thrusts, is required to address this issue.

Conclusions

(1) Samples of surface glacier ice on Midtre Lovénbreen plot on a line with a slope of 6.9 on a bivariate co-isotopic plot (δD–$\delta^{18}O$). This line closely approximates the local meteoric line obtained from snow samples (6.4) and that obtained from monthly precipitation data collected by IAEA at sea level in Ny-Ålesund (7.2). This suggests that the surface ice on Midtre Lovénbreen has not undergone significant isotopic modification during the transformation from snow to glacier ice.

(2) There is clear evidence of isotopic homogenization during the transformation from snow to glacier ice. When plotted on a bivariate co-isotopic diagram, the extreme values of δD and $\delta^{18}O$ present in the snow are lost from the surface glacier ice.

(3) There is no evidence that the snow becomes isotopically lighter with altitude on the glacier. There is no clear indication why this should be, but it is possible that the samples collected may either represent multiple snowfall events or that isotopic values at depth in the snowpack are affected by meltwater percolation during firnification. A larger data set addressing temporal and spatial differences in the isotopic content of the original snow is required to determine which of these processes is responsible.

(4) When plotted on a bivariate co-isotopic diagram (δD–$\delta^{18}O$), the slopes for snow and unmodified glacier ice (6.4 and 6.9, respectively) are less steep than those for the basal ice layer and transverse ice layers on the ice surface (7.6 and 7.7, respectively). However, the difference in the slope of these lines is not statistically significant at the sample size used in this study.

(5) The results highlight some of the difficulties in using δD–$\delta^{18}O$ relationships to define the origin of glacier structures on polythermal glaciers. Difficulties centre on the relatively small number of samples analysed, the possibility of diagenetic fractionation and uncertainties concerning the relative importance of open-system and closed-system freezing conditions. As a consequence, it is not yet possible to apply these isotopic techniques to determine the genesis of ice facies of unknown origins.

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