Stable Isotope Geochemistry: A Tribute to Samuel Epstein
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Isotopic changes during snow metamorphism

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Abstract—Mechanisms of metamorphism of snow were investigated by examining the deuterium and oxygen-18 content of snowpacks deposited in Fairbanks, Alaska. This location was chosen because of the very large thermal gradient in the snowpack. Experiments were carried out for nine years by sampling two or more adjacent snowpack regimes. The first snowpack was deposited on bare ground (soil-pack). The second snowpack was deposited on ground that was covered by a plastic sheet to prevent water-vapor transfer from the ground to the snow above it (plastic-pack). The third snowpack was deposited on a nearby wooden table (table-pack). The table-pack did not experience the thermal gradient between top and bottom that was developed by the first two, and is used as a sample of the original unaltered snow. All sampling was carried out in the spring, after snow metamorphism had occurred, but before melting of the snow. Although all of the snowpacks lying on the soil or plastic developed a typical depth-hoar profile, none of the table-packs showed evidence of metamorphism associated with the development of depth-hoar. All of the snowpacks displayed variations of isotopic composition with depth associated with isotopic differences between individual precipitation events. In all of the vertical sections, the bottommost 5 to 20 cm of the soil-packs was enriched in deuterium relative to the table-packs, indicating that water molecules depleted in deuterium (and oxygen-18) had been removed from this portion of the snowpack. This isotopically light material appears to have been incorporated in the snow immediately above this bottom portion of the soil-pack, which was lower in δD and $\delta^{18}O$ relative to the corresponding portion of the table-pack. Isotopic analyses of the water vapor that left the soil and was condensed as ice under the plastic sheet have δD and $\delta^{18}O$ values more positive than any of the snow. The transfer of water from the soil to the immediately overlying 10–20 cm of snow played a part in changing the isotopic composition of the natural soil-packs. The slopes of the lines joining the bottom samples of the table-pack and soil-pack in a δD vs. $\delta^{18}O$ plot ranged from 3 to 4.3, in contrast to the slopes found in δD vs. $\delta^{18}O$ plots of all of the samples from the table-packs, which ranged from 7.4 to 8.8, characteristic of the Meteoric Water Line. Slopes of 2 are characteristic of diffusive isotope separation processes, and the low slopes found in the bottom samples indicate the role of diffusion transport of water both from the soil to the overlying snowpack, and also within the snow. Higher slopes were found in the remainder of the samples and are the result of fractionation due to change-of-state processes, as well as the exchange of water vapor present in the atmosphere above the snow with water vapor in the snowpack.

INTRODUCTION

THE ISOTOPIC CHANGES that accompany the metamorphism of snow under natural conditions have received little attention. EPSTEIN *et al.* (1965) observed that the ^{18}O content of depth-hoar was higher than that of the remainder of the snowpack. MACPHERSON and KRAUSE (1967) determined that percolating melt water homogenized the $\delta^{18}O$ of the snow. JUDY *et al.* (1970) showed that changes in the δD of the snow occurred as the snow pack aged, resulting in a greatly reduced variability of δD from the original snowfalls.

Because deuterium and oxygen-18 are fractionated in slightly different ways during processes that are mass dependent (for example, diffusion), we hoped that a study of both isotopes in natural snowpacks might yield information on the mechanisms of depth-hoar formation. Depth hoar constitutes a weak layer at the bottom of the snowpack and has long been studied for its role in producing avalanches on steep slopes (BADER *et al.*, 1939; YOSIDA *et al.*, 1955). Depth hoar is also responsible for the formation of recognizable annual layers in the dry snow strata of polar ice sheets (SORGE, 1935; BENSON, 1962).

The study was carried out on samples of the seasonal snowpack collected in Fairbanks, Alaska by one of us (C.B.). The site at Fairbanks is ideal for such a study because the snowpack lasts for a long time (150–200 days) at temperatures well below freezing (−40°C is common) and has a very large thermal gradient (1°C/cm is common) established between the ground at its base and the air above it. The resulting depth hoar has exceptionally low density (0.19 to 0.20 g/cm^3 compared with 0.28 to 0.30 g/cm^3 in alpine regions), and it can spread upwards through the entire snowpack before winter is over (TRABANT and BENSON, 1972; STURM, 1989).

EXPERIMENTAL

All collections were made at the end of winter in March and early April, before melting of the snow began. Snow samples were taken with a set of stainless steel sample tubes inserted horizontally into the vertical walls of pits dug through the snow pack. The 500 cm^3 samples were weighed and density values were calculated. The samples were emptied into plastic bags in the field and sealed until they could be melted and poured into tightly sealed bottles to minimize evaporation of the melted snow. They were then sent to the U.S. Geological Survey laboratories in Denver for δD and $\delta^{18}O$ analysis.

The δD analyses were made by reacting 5 μl samples of water with hot uranium metal. The hydrogen gas produced by this reaction was analyzed for deuterium using a double collector Nier-type mass spectrometer of 15 cm radius. All results are expressed in delta units relative to Vienna Standard Mean Ocean Water (V-SMOW), where

$$\delta D = \frac{\left(\frac{D}{H}\right)_{\text{sample}} - \left(\frac{D}{H}\right)_{\text{standard}}}{\left(\frac{D}{H}\right)_{\text{standard}}} \times 1000$$

and were normalized to zero for Vienna Standard Mean Ocean water (V-SMOW), and −428 for Standard Light Antarctic Precipitation (SLAP). The deuterium analyses were carried out in replicate, with a two sigma of 0.8 per mil.

The $\delta^{18}O$ analysis were made by equilibrating a 5 ml sample of water with CO_2, purifying the CO_2, and analyzing the purified gas on a double collecting Nier-type mass spectrometer. After making appropriate corrections the results are given in delta units relative to V-SMOW, with a two sigma of 0.1 per mil. The results have been normalized to zero for V-SMOW and −55.5 for SLAP:

$$\delta^{18}O = \frac{\left(\frac{^{18}O}{^{16}O}\right)_{\text{sample}} - \left(\frac{^{18}O}{^{16}O}\right)_{\text{standard}}}{\left(\frac{^{18}O}{^{16}O}\right)_{\text{standard}}} \times 1000$$

The analytical results are presented in Table 1.

COLLECTION PROTOCOL

Experiments were carried out by sampling two or more adjacent snowpack regimes as described by TRABANT and BENSON (1972):

(1) The snowpack resting on bare ground is referred to as "soil-pack."

(2) The snowpack on ground that was covered with a plastic sheet to prevent water vapor transfer from the ground to the snow above it is referred to as "plastic-pack."

(3) An adjacent pack resting on a wooden table in a wind-sheltered area in a spruce forest is referred to as "table-pack." This pack did not experience the thermal gradient between top and bottom developed by the first two packs, because it had cold air below as well as above. The fundamental assumption is that negligible vapor flux occurs within the snowpack resting on the table, and there is no possibility of vapor transfer into it from the soil. Thus, the table snowpack was treated as a control each year and the other snowpacks were compared to it.

All sampling was carried out before melting occurred. In 1966, 1972, 1974, 1975, 1976, and 1982, only the soil-packs and table-packs were sampled and were analyzed for deuterium. In addition to deuterium, the 1974, 1982, 1985 and 1986 samples were also analyzed for oxygen-18. In 1985 the sampling protocol was extended to include the plastic-pack, in addition to the soil-pack and table-packs. Samples of ice from the underside of the plastic sheet were collected in 1975, 1985, and 1986.

RESULTS

Physical changes in the snowpack

In most years we observed that the table-packs were thinner than the adjacent soil and plastic-packs (Figs. 1 through 9). The table-packs underwent some horizontal creep, resulting in the formation of snow cornices at the outer edges of the table. Although part of the reduction in thickness of the table pack was due to this horizontal movement, the depth hoar is mechanically stronger than fine-grained compact snow against a static force, even though it is extremely fragile against a dynamic force (KOJIMA, 1956; AKITAYA, 1974). Thus, depth hoar on the ground does not settle under the load of overlying snow as fast as the fine-grained compact snow on the tables. The effect of this on our experimental setup was discussed by STURM (1989). The different compaction rate of the table-packs complicates comparisons between similar portions of the table-pack and of the other two snowpacks. The further up from the bottom that such comparisons are made, the more uncertain that any chosen horizontal horizon in the table-pack is equivalent to that in the soil-pack or plastic-pack. Therefore we have restricted comparisons between the table-pack and the other packs to the lowest and next-to-lowest sampled horizons.

Although all of the soil and plastic-snowpacks developed a typical depth-hoar profile, none of the table-packs developed depth-hoar, emphasizing the importance of a thermal gradient in the snowpack for depth-hoar development to occur. Every year that that we have compared the table-packs and soil-packs, we have found that the table-pack had a higher density at the bottom but a lower density at the top. A summary of the density distribution of the soil-packs and table-packs during 11 winters between 1966 and 1987 is shown in normalized form in Fig. 16 (from STURM, 1989).

Figures 1–9 are plots of δD vs. depth in the snow pack for all collections, while Figs. 10–13 are plots of δD vs. $\delta^{18}O$ for 1974, 1982, 1985, and 1986. Because the lowest sample from each of the soil-packs had suffered large changes in isotopic composition from soil moisture, as discussed below, the least mean squares solution to the data sets in Figs. 10–13 have not included data for this lowest soil-pack sample.

Isotopic composition of snow

All of the snowpacks display variations of isotopic composition with depth associated with isotopic differences between individual precipitation events. During the winter, as the snowpacks that experienced a thermal gradient (soil-pack, plastic-pack) aged, physical changes occured that modified these initial isotopic differences (JUDY *et al.*, 1970).

Soil moisture

Measurement of soil moisture in Fairbanks over many years has shown that the moisture content at the beginning of snow accumulation is about 30% by dry weight and is reduced to less than 10% by March (TRABANT and BENSON, 1972; STURM, 1989). Calculations show that about 1 cm of water equivalent is transferred from the soil to the overlying snowpack during the winter. This is about 10% of the total snowpack.

Ice that accumulated on the underside of the plastic-ground cover was collected in 1975, 1985, and 1986. In all three cases, the isotopic composition of this water was enriched in the heavier isotopes as compared to the snowpacks. We ascribe this to the fact that the soil moisture is a mixture of precipitation that fell as summer rain as well as winter snow. The summer rain can be expected to be enriched in the heavier isotopes, because the condensation temperature of the summer rain is higher than that of the winter snow, resulting in less loss of the heavier isotopic molecules from the cloud system that is undergoing precipitation (FRIEDMAN *et al.*, 1964).

This soil moisture was important in modifying the isotopic composition of the bottom of the soil-pack. However, as will be discussed later, the bottom soil-pack samples are not simply soil moisture, nor mixtures of soil moisture and pre-existing snow, but can be modelled by mixtures of soil moisture and snow that has undergone heavy-isotope enrichment after deposition.

δD vs. depth

In all of the δD vs. depth plots (Figs. 1–9) the one or two lowest samples from the base of the soil-packs are higher in δD relative to the corresponding table-pack. Inasmuch as the lower section of the soil-pack had a lower density than the corresponding section of the table-pack, this indicates that material that was depleted in deuterium (and oxygen-18) had been removed from this portion of the soil-pack. This heavy-isotope–depleted material appears to have been incorporated in the snow immediately above this bottom portion of the soil-pack, the 10 to 20 cm layer, which is depleted in deuterium and oxygen-18 relative to the corresponding portion of the table-pack.

δD vs. δ^{18}O

On a δD-δ^{18}O plot of the 1974, 1982, and 1986 data (Figs. 10, 11, and 13) the table- and soil-pack data plot very close to each other. However, in a similar plot of the 1985 data (Fig. 12) the soil-pack data displays a small shift relative to the table-pack data.

The relation between the δD and δ^{18}O values of the basal samples from the soil-pack and from the table-pack from the 1974, 1982, 1985, and 1986 collections is shown in Fig. 14. We have joined the plotted values of the bottom samples from the soil and from the table-packs, inferring that the samples from the table-pack represent the starting snow unaltered by processes that originated in the soil, and that the sample from the base of the soil-pack has been changed isotopically by such processes. In all of the four cases plotted, the basal soil-pack sample had been enriched by about 15 per mil in deuterium and about 3.3 per mil in oxygen-18.

On a δD vs. δ^{18}O plot, the slopes of the lines that join the values of the basal samples in the table- and soil-pack range from 3 to 4.3, in contrast to the slopes found for all of the remaining samples (see Figs. 10–13), which range from 7.4 to 8.8. These higher slopes are characteristic of isotope separation processes that depend upon the change of state between solid and vapor. The slope of the line joining the bottom samples from the 1986 *plastic*-pack and *table*-packs is 2.8. As will be discussed later, slopes of 2 to 5 are characteristic of diffusive isotope separation processes and indicate the role of diffusive transport of water both from the soil to the overlying snowpack, as well as within the lower portion of the snowpack itself.

A δD vs. δ^{18}O plot of the samples collected from the interval 10–20 cm for these same years is shown in Fig. 15. The slopes of the lines joining the corresponding soil and table points for each year range from 7.4 to 13, indicating that changes in isotopic composition of samples from this level in the snow-

Table 1. Isotopic data on snow packs

Year	Location	Depth (cm)	$\delta^{18}O$	δD
1966[1]	soil	0–10	—[2]	−158
		10–20	—	−165
		20–30	—	−205
		30–41	—	−160
		41–43	—	−174
		43–49	—	−174
		49–53	—	−150
		53–60	—	−153
		60–65	—	−172
		65–73	—	−225
	table	0–4	—	−167
		4–9	—	−162
		9–16	—	−171
		16–26	—	−207
		26–33	—	−156
		33–40	—	−167
		50–56	—	−164
		56–60	—	−169
		60–67	—	−194
1972[3]	soil	2–5	—	−166
		5–10	—	−179
		10–15	—	−185
		15–20	—	−186
		20–25	—	−213
		25–30	—	−202
		30–35	—	−198
		35–40	—	−165
		40–43	—	−181
		43–48	—	−195
		48–55	—	−179
		55–60	—	−146
		60–65	—	−218
	table	1–5	—	−175
		5–7	—	−183
		7–10	—	−179
		10–15	—	−209
		15–20	—	−224
		20–25	—	−196
		25–30	—	−191
		29–33	—	−198
		34–40	—	−202
		40–45	—	−170
		45–50	—	−150
		50–56	—	−182
1973[4]	soil	0–4	—	−146
		5–10	—	−156
		10–14	—	−160
		14–18	—	−159
		18–20	—	−147
		20–24	—	−149
		24–28	—	−165
		28–31	—	−191
		31–35	—	−204
		35–39	—	−214
		39–43	—	−204
		43–47	—	−176
1973[4]	table	0–4	—	−153
		4–8	—	−174
		8–10.5	—	−179
		10.5–13.5	—	−148
		13.5–17.0	—	−117
		17.0–20.5	—	−115
		20.5–23.0	—	−138
		23.0–26.0	—	−184
		26.0–29.5	—	−196
		29.5–32.8	—	−202
		32.8–35.9	—	−215
		39–42	—	−175
		42–46	—	−183
1974[5]	soil	0–5	−19.0	−154
		0–5	−19.6	−154
		5–10	−21.9	−157
		10–13	−20.8	−154
		13–15	−22.1	−154
		15–20	−23.3	−161
		20–25	−26.1	−190
		25–30	−28.3	−204
		30–35	−28.1	−206
		35–40	−31.6	−245
		45–50	−29.0	−220
	table	0–5	−24.4	−184
		0–6	−24.8	−179
		0–5	−24.3	−177
		5–10	−22.5	−166
		10–15	−20.0	−144
		15–20	−24.1	−178
		20–25	−28.1	−208
		25–30	−28.0	−203
		30–35	−29.1	−218
		35–40	−31.0	−250
		40–45	−29.0	−222
		45–50	−26.1	−201
1975[6]	soil	0–5	—	−172
		5–8	—	−185
		8–10	—	−192
		10–14	—	−197
		14–18	—	−200
		18–22	—	−200
		22–25	—	−199
		25–29	—	−198
		29–34	—	−201
		34–38	—	−206
		35–41	—	−214
		42–48	—	−261
		48–54	—	−137
	table	0.5–6.5	—	−201
		7–13	—	−202
		13–19	—	−199
		19–25	—	−195
		25–31	—	−203
		32–38	—	−249
		38–44	—	−191
		44–56	—	−136
		48.5–54.5	—	−177
	under plastic		—	−147

Table 1. (Continued)

Year	Location	Depth (cm)	$\delta^{18}O$	δD	Year	Location	Depth (cm)	$\delta^{18}O$	δD
1976[7]	soil	0–4	—	−144	1985[9]	table	0–6	−22.6	−169
		4–7	—	−166			4–10	−25.6	−183
		1.5–4.5	—	−140			11–17	−18.0	−134
		4–7	—	−158			18–24	−19.7	−147
		7–10	—	−167			25–31	−21.6	−164
		10–13	—	−171			34–40	−25.9	−194
		13–16	—	−165			41–47	−26.6	−204
		16–19	—	−163			49–55	−23.7	−183
		19–22	—	−170			56–62	−21.9	−170
		22–25	—	−188			64–70	−23.0	−177
		25–28	—	−199		plastic	1–7	−21.2	−170
		28.5–31.5	—	−193			5–11	−24.3	−180
		35.5–35.5	—	−217			10–16	−20.6	−151
	table	0–3	—	−164			16–22	−18.6	−133
		0–3.5	—	−162			21–27	−20.6	−152
		3.5–7	—	−172			28–34	−23.2	−174
		7–10	—	−161			34–40	−25.9	−194
		10–13	—	−166			40–46	−26.4	−201
		19–22	—	−191			46–52	−25.7	−200
		22–25	—	−216			52–58	−23.9	−181
		25–28	—	−164			58–64	−25.1	−191
1982[8]	soil	1–7	−21.5	−184			64–70	−18.5	−139
		9–15	−25.8	−211			69–75	−21.0	−156
		17–23	−26.4	−214			74–80	−22.6	−172
		23–29	−25.0	−193			1–7		−163
		27–33	−23.8	−182		under plastic		−13.4	−117
		33–39	−24.0	−189				—	−118
		32–38	−21.4	−174				−12.9	−115
	table	1–7	−24.8	−197	1986[10]	soil	0–6	−21.8	−176
		6–12	−26.9	−211			7–13	−24.2	−181
		11–17	−27.2	−213			14–20	−19.5	−142
		16–22	−24.8	−192			18–24	−19.5	−138
		23–29	−23.5	−184			24–25	−21.3	−168
		28–34	−25.4	−205			26–29	−26.0	−198
1985[9]	soil	2–8	−18.0	−155		table	1–7	−25.1	−190
		9–15	−23.6	−183			4–10	−21.2	−159
		14–20	−21.4	−162			10–16	−20.7	−155
		21–27	−18.3	−144			12–18	−23.0	−179
		27–33	−20.6	−164		plastic	1–7	−22.2	−182
		33–39	−23.2	−182			7–13	−22.5	−174
		38–44	−25.6	−207			9–15	−21.7	−159
		45–51	−26.2	−209			14–20	−19.7	−145
		52–58	−24.5	−198			16–22	−21.0	−156
		58–64	−24.4	−200			24–25	−22.0	−171
		64–70	−19.3	−156			25–28	−24.6	−198
		69–75	−21.3	−177		under plastic		−17.7	−145
		71–77	−22.5	−185					

[1] Samples collected March 22, 1966.
[2] Dash indicates no measurements were made.
[3] Samples collected March 27, 1972.
[4] Samples collected March 25, 1973.
[5] Samples collected March 16, 1974.
[6] Samples collected March 27, 1975.
[7] Samples collected April 3, 1976.
[8] Samples collected March 12, 1982.
[9] Samples collected March 18, 1985.
[10] Samples collected March 19, 1986.

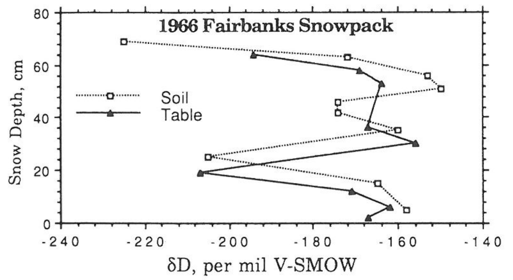

FIG. 1. A plot of δD vs. vertical position in the soil-pack and table pack for the 1966 snowpack sampled in the spring before melting occurred.

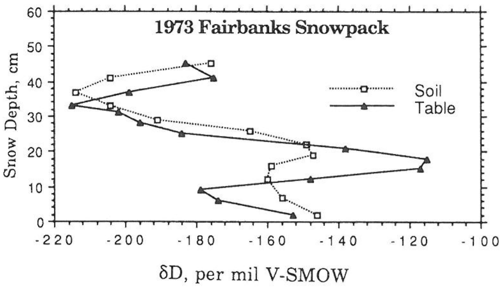

FIG. 3. A plot similar to Fig. 1, but for the 1973 snowpack.

pack was determined predominately by vapor-pressure fractionation, or by exchange with atmospheric water vapor, as will be discussed later. The δD vs. $\delta^{18}O$ slope for the 10–20 cm interval for 1986, but for the plastic-pack and table-pack samples, is 12, showing that similar processes have occurred in these levels of both the plastic-pack and soil-pack.

DISCUSSION

The isotopic data from the 1985 and 1986 plastic-packs resembles that from the table-packs, and differs from the soil-packs. This indicates that the plastic sheet intercepted an agent that would normally act on the snowpack above the soil. The plastic sheet prevented transfer of moisture from the soil to the snow above it. The moisture condensed under the plastic sheet was sampled in 1975, 1985, and 1986. In all cases this condensed moisture, which normally would have moved into the snow above it in the soil-pack, is very enriched in deuterium and oxygen-18 relative to the snow. In Figs. 12 and 13 we have plotted the data obtained from samples of this under-plastic condensate. From these figures it would appear that the bottom soil-pack samples are a simple mixture of this condensate with the lowest *plastic* sample, but not with the lowest *table* sample. Therefore, the agents that acted to change the isotopic composition of the base of the soil-pack are not only dilution or mixing with the soil-derived moisture, but fractionation associated with transport of water within the snowpack itself.

This transport can occur by several mechanisms. (1) Surface diffusion along the snow crystal surfaces and transfer of material from crystal-to-crystal at the contact between crystals can be expected to fractionate both deuterium and oxygen-18. (2) Vapor transport from the bottom crystals to the top crystals, along the vapor pressure gradient generated by the temperature gradient in the snow pack, will result in an isotopic fractionation that is caused by two processes. The first process is the change of state from solid to vapor, while the second is molecular diffusion during vapor transport.

Fractionation during change of state

The vapor pressure of the lighter isotopic molecule ($H_2{}^{16}O$) is higher than that of the heavier isotopic molecules (HDO and $H_2{}^{18}O$); therefore, the vapor will be enriched in the light molecules relative to the solid in equilibrium with it. EPSTEIN *et al.* (1965) observed that depth hoar was enriched in

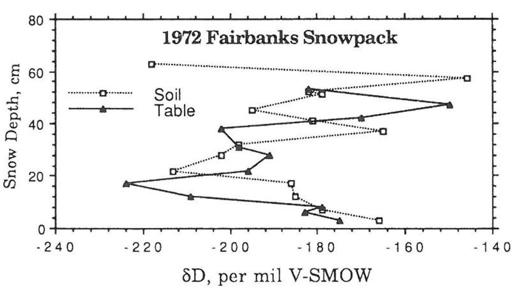

FIG. 2. A plot similar to Fig. 1, but for the 1972 snowpack.

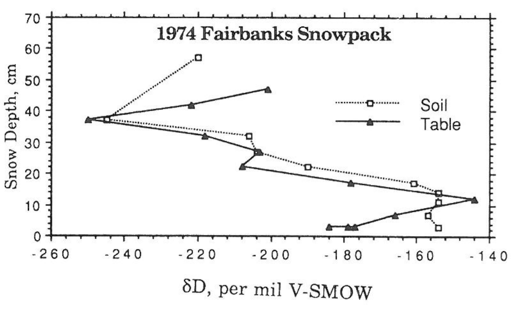

FIG. 4. A plot similar to Fig. 1, but for the 1974 snowpack.

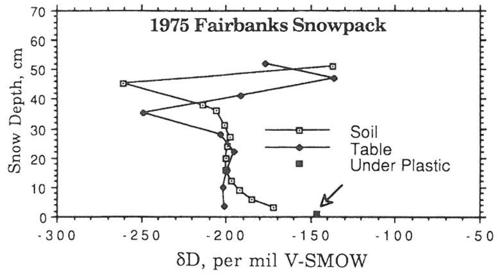

FIG. 5. A plot similar to Fig. 1, but for the 1975 snowpack. The δD of the frost condensed under a plastic sheet covering the soil is also shown.

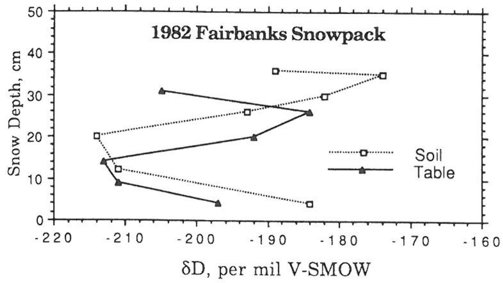

FIG. 7. A plot similar to Fig. 1, but for the 1982 snowpack.

^{18}O and suggested that this was the result of partial recondensation of vapor generated within the depth hoar accompanied by the escape of residual vapor impoverished in ^{18}O. The difference in vapor pressure of the isotopic molecules results in the vapor being lower in δD by 122 per mil and in $\delta^{18}O$ by 11.8 per mil at 0°C, and by 160 and 13.4, respectively, at −20°C (FRIEDMAN and O'NEIL, 1977). The slope of δD vs. $\delta^{18}O$ for metamorphosed snowpack samples should be between 10 and 12 (the ratio of the vapor pressures of the isotopic molecules) if this process was the dominant transport agent. The solid-vapor fractionation due to the vapor pressure difference between the isotopic molecules can, in the steady state, only be effective in isotopic separation if one or both of the following processes occur.

(1) First is fractionation during sublimation of snow. In order for this process to be effective in isotope separation, the rate of self-diffusion in the solid must be rapid compared to the rate of molecular diffusion in the vapor. In other words, the solid must "mix" so that the surface layer of the solid undergoing sublimation will not be unduly enriched in the heavier isotopic species due to the preferential loss of the lighter isotopic species to the vapor. This would result in sublimation occurring "layer-by-layer" with no resulting isotopic fractionation. Because self-diffusion in ice is about 1/1000 as rapid as molecular diffusion in the vapor, the amount of isotopic separation due to vapor pressure differences between the isotopic molecules will be greatly reduced from the values of approximately 130 per mil for deuterium and 12 per mil for oxygen-18 that would to be expected if the molecules in the solid were well mixed.

(2) The second process is the condensation of vapor produced in the snowpack on growing ice crystals. If partial condensation of rising vapor occurs at each growing crystal, then the crystals receiving the condensate would be enriched in the molecules having the lowest vapor pressure (the heavier molecules). Under our extreme conditions, virtually all of the ice present at the base of the Alaskan snowpack passes from solid to vapor and back into the solid phase during the winter, according to TRABANT and BENSON (1972) and STURM (1989). These authors documented this by showing that the number of grains decreased from about 400 per cm³ at the beginning of the winter to about 40 at the end, and these few crystals grew to much larger sizes (up to 1 cm) than the original

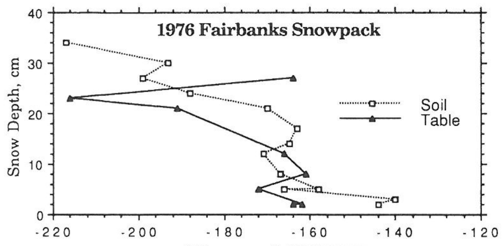

FIG. 6. A plot similar to Fig. 1, but for the 1976 snowpack.

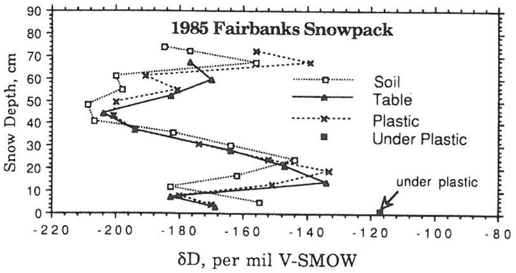

FIG. 8. A plot of δD vs. vertical position in the soil-pack, table-pack, and plastic-pack for the 1985 snowpack sampled in the spring before melting occurred. The δD of the frost condensed under the plastic sheet covering the soil is also shown.

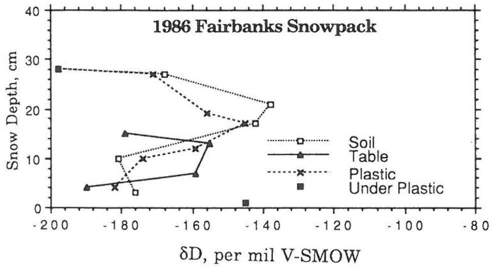

FIG. 9. A plot similar to Fig. 8, but for the 1986 snowpack.

crystals (<1 mm). This order-of-magnitude decrease in the number of grains means that only one in ten grew to become large depth-hoar crystals by sublimation while the remainder vanished. In effect, almost all of the snow in the depth hoar will have sublimed, passed into the vapor state, and then condensed on a relatively few ice crystals. Even though this change-of-state process operating at reduced efficiency results in only a small isotopic fractionation, the cascading of many such change-of-state events, the "hand-to-hand" transfer mechanism of YOSIDA *et al.* (1955), as the vapor travels vertically through the snowpack can result in relatively large isotopic changes, with slopes of the δD vs. $\delta^{18}O$ changes having values in the range of 10 to 12.

Atmospheric water vapor

Air moved in-and-out of the snowpack due to changes in temperature and atmospheric pressure (TRABANT and BENSON, 1972; GJESSING, 1977). Atmospheric water vapor will have an isotopic composition that is approximately in equilibrium with that of the "average" snow, and this water vapor can exchange with vapor produced in the snow. This exchange will act to maintain the δD-$\delta^{18}O$ slope at values of about eight, similar to that in the original snow, and will be more effective in the upper, rather than lower, portion of the snowpack, inasmuch as the upper portion experiences the largest amount of air exchange.

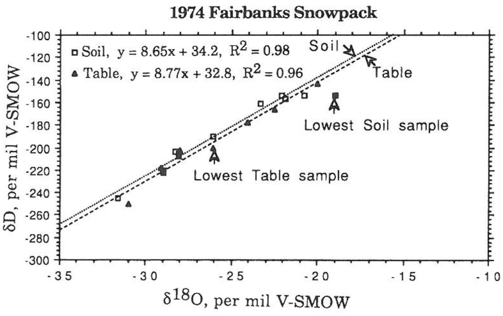

FIG. 10. A plot of δD vs. $\delta^{18}O$ for the soil-pack and table-pack for the 1974 snowpack sampled in the spring before melting occurred. The straight lines are the linear least mean square solutions to the soil-pack and to the table-pack data. The data from the lowest soil sample were not included in the calculation of the linear least mean square solution to the soil-pack data.

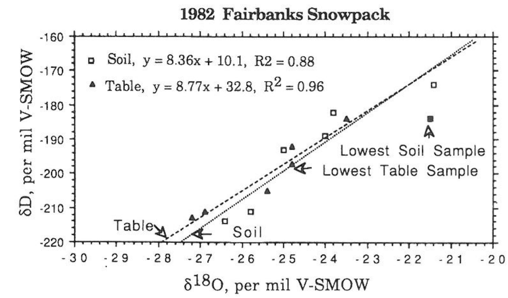

FIG. 11. A plot similar to Fig. 10, but for the 1982 snowpack.

Isotopic fractionation due to molecular diffusion

Molecular diffusion of water along the surfaces of crystals can result in isotopic fractionation. This diffusive separation is a function of the ratio of the relative masses of the diffusing molecules, and the separation factor ϵ is equal to $1 - \alpha$, where α is the fractionation factor defined as $\alpha = \frac{R_{\text{initial phase}}}{R_{\text{final phase}}}$, where

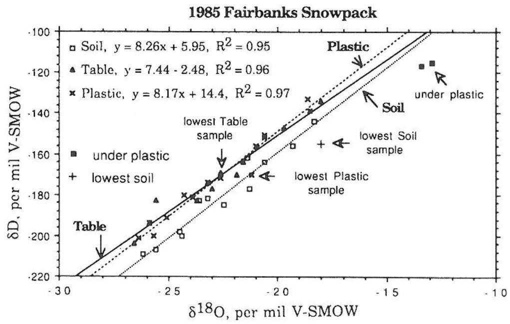

FIG. 12. A plot of δD vs. $\delta^{18}O$ for the soil-pack, table-pack, and plastic-pack for the 1985 snowpack sampled in the spring before melting occurred. The isotopic composition of the ice collected from the underside of the plastic sheet is also plotted. The straight lines are the linear least mean square solutions to the soil-pack, table pack, and plastic-pack data. The data from the lowest soil-pack and the lowest plastic-pack samples were not included in the calculation of the linear least mean square solution to the corresponding soil-pack and plastic-pack data.

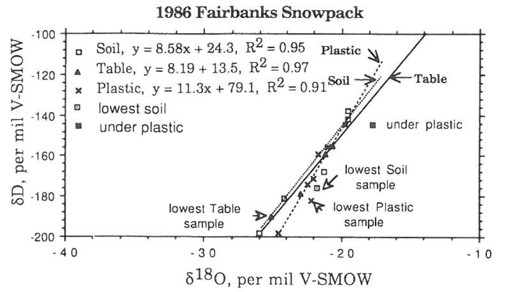

FIG. 13. A plot of δD vs. δ^{18}O for the soil-pack, table-pack, and plastic-pack for the 1986 snowpack sampled in the spring before melting occurred. The isotopic composition of the ice collected from the underside of the plastic sheet is also plotted. The straight lines are the linear least mean square solutions for the soil-pack, table-pack, and plastic-pack data. The data from the lowest soil-pack and the lowest plastic-pack samples were not included in the calculation of the linear least mean square solution to the corresponding soil-pack and plastic-pack data.

R is the ratio of the concentration of heavy to light isotopes:

$$\epsilon_{\text{deuterium}} = 1 - \alpha_{\text{D}} = 1 - \sqrt{\frac{19}{18}} = .0274$$

$$\epsilon_{\text{oxygen-18}} = 1 - \alpha_{\text{O}} = 1 - \sqrt{\frac{20}{18}} = .0540.$$

The ratio of the two separation factors given above is 1.97 and will be equal to the slope of the δD vs. δ^{18}O plot of the samples undergoing such a diffusive separation.

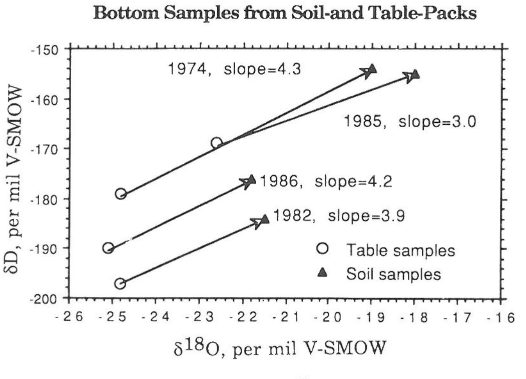

FIG. 14. A plot of the δD vs. δ^{18}O for the samples from the bottom of the soil-pack and table-pack of the 1974, 1982, 1985, and 1986 snowpacks. Lines are drawn to connect the corresponding soil-pack and table-pack samples from each year. The arrows indicate the direction of the changes, from the original snow in the table-packs to the metamorphosed snow in the soil-packs. The slopes of these lines are also shown.

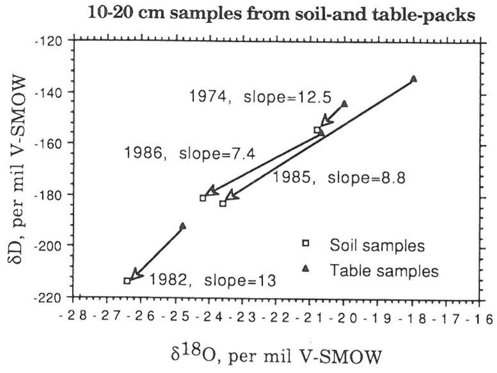

FIG. 15. A plot similar to Fig. 14, but for samples from 10–20 cm above the bottom of the snowpacks.

Molecular diffusion of water vapor through the air during the transport of vapor from bottom to top of the snowpack will also result in fractionation. EHHALT and KNOTT (1964) found that the separation factors ϵ are given by

$$\epsilon_{\text{deuterium}} = 1 - \alpha_{\text{D}}$$

$$= 1 - \sqrt{\frac{19(18 + 29)}{18(19 + 29)}} = .01664$$

$$\epsilon_{\text{oxygen-18}} = 1 - \alpha_{\text{O}}$$

$$= 1 - \sqrt{\frac{20(18 + 29)}{18(20 + 29)}} = .03236.$$

The ratio of the two separation factors is 1.95.

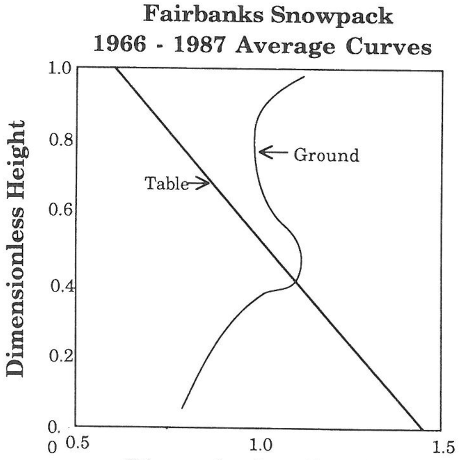

FIG. 16. A plot, in normalized form (from STURM, 1989), of the density distribution of the soil and table-packs during 11 winters between 1966 and 1987.

Experimental determination of the fractionation factors for diffusion of water vapor in air was determined by MERLIVAT (1970) and STEWART (1974), and gave separation factors of .024 for deuterium and .0289 for oxygen-18. The ratio of these separation factors is 1.20.

EHHALT and KNOTT (1964) and CRAIG and GORDON (1965) predicted values of between three and five for the case of evaporation from a liquid, where similar vapor pressure differences and molecular transport phenomena can be expected, but where mixing in the liquid phase is more rapid than in the solid snow.

In a laboratory study of snow metamorphism, SOMMERFELD *et al.* (this volume) observed that the slope of the δD vs. $\delta^{18}O$ of all of the samples from the experimental snowpack that had been exposed to a temperature gradient was 3.0. In this experiment, the snowpack was not in contact with normal atmospheric water vapor, but was contained in a freezer, with the result that vapor was continuously being removed from the snow and transferred to the freezer coils. This one-directional flow of water vapor resulted in the experimental snowpack having a slope of 3.0, in contrast to the values of 10 to 13 found for the natural snowpacks. In another laboratory study of snow metamorphism, SOMMERFELD *et al.* (1986) found that the slope varied from 4.0 to 4.5, again much lower than that of the natural snowpack. In this later study the snow was sealed in plastic bags and isolated from the atmosphere.

SOMMERFELD *et al.* (1986) found that the effective diffusion coefficient of water vapor in snowpacks exposed to a temperature gradient was approximately twice that of water vapor in air. This result indicates that hand-to-hand water transport along the surface of snow crystals and from crystal-to-crystal is an important factor in vapor transport in the snow pack.

Relative importance of soil-added water and heat transport in the metamorphism of snow

In both 1985 and 1986 the plastic-packs experienced about the same degree of metamorphism as did the soil-packs. Inasmuch as water from the underlying soil could not enter the plastic-packs, water from the underlying soil was not *directly* involved in the metamorphism of these packs. The major driving-force in this metamorphic change was heat transfer from the soil to the snow. In a natural snowpack, in addition to conduction, an important agent in this heat transfer is the latent heat transfer associated with the water vapor as it moves from the ground and condenses in the snowpack. In the experiment where a plastic sheet prevented the direct transfer of vapor from the ground to the snow, the vapor from the soil was condensed on the underside of the plastic, transferring heat to the plastic, and thereby to the snow above. In this case heat transport by water vapor was still effective in inducing the recrystallization of the snow, although water molecules from the soil did not reach the snow.

The primary factor in snow metamorphism is the addition of heat from the ground to the overlying snowpack. This heat is transferred along a thermal gradient from the warm ground to the cooler snow. The intensity of this gradient is determined by a number of climatic factors, including the timing of the first permanent snow cover, the temperature at which each snowfall occurs, and the intensity and duration of cold events during the winter.

This temperature gradient acts to transfer both heat and water from the external environment of the soil to the snowpack. In addition, it acts to redistribute water within the snowpack by diffusive transport. Each of these processes acts somewhat differently on the isotopic water species and allows us to separate the processes.

CONCLUSIONS

1. A thermal gradient within the snowpack is necessary for the development of depth hoar. This is shown by the development of depth hoar in the plastic-pack, which developed a thermal gradient, but not in the table-pack where no thermal gradient existed.
2. Transfer of water from the underlying soil to the snowpack is not necessary for depth-hoar formation. This was shown by the development of depth hoar in the plastic-pack, where the plastic sheet prevented water from the soil from reaching the overlying snow.
3. Although atmospheric water vapor interacts with the natural snowpack, this interaction is not necessary for depth-hoar development to occur. The experimental snowpacks, particularly in the experiments of SOMMERFELD *et al.* (1986), were not in contact with atmospheric water vapor but did develop depth hoar.
4. Atmospheric water vapor interacts with the natural snowpack and is responsible for some isotopic alteration of the snowpack.
5. Fractional condensation of water vapor (produced by sublimation of the snow), on growing snow crystals as the vapor moves vertically through the snowpack, results in isotopic fractionation.
6. The efficiency of the Rayleigh process during

sublimation is reduced due to the low (diffusive) rate of mixing of water molecules within the snow crystals, and isotopic fractionation during the sublimation of snow is therefore not an important process.

7. Transport of water vapor by diffusion is an important process. However, we do not know the relative importance of diffusion through the intergranular atmosphere as compared to diffusion along a liquid water film as described by Dash (1990).

8. In the lowest portion of the natural snowpack, diffusive processes are most important in mass-transfer, while further up in the snowpack change-of-state processes and vapor advection usually predominate. These differences may be due to the fact that the amount of advected atmospheric water vapor in the snowpack decreases in proportion to the depth below the snow-atmosphere interface.

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