HOKKAIDO UNIVERSITY

Title	Isotope studies of a snow layer in a temperate region : isotope balance and temporal change
Author(s)	Nakawo, Masayoshi; Hashimoto, Shigemasa; Ageta, Yutaka et al.
Description	International Symposium on Physics of Ice Core Records. Shikotsukohan, Hokkaido, Japan, September 14-17, 1998.
Citation	Physics of Ice Core Records, 141-151
Issue Date	2000
Doc URL	https://hdl.handle.net/2115/32466
Type	conference paper
File Information	P141-151.pdf

Hokkaido University Collection of Scholarly and Academic Papers : HUSCAP

Isotope studies of a snow layer in a temperate region - isotope balance and temporal change

Masayoshi Nakawo*, Shigemasa Hashimoto*, Yutaka Ageta* and Nobuo Miyazaki**

*Institute for Hydrospheric-Atmospheric Sciences, Nagoya University, Furo-cho, Nagoya 464-8601, JAPAN
**Climate Engineering Co., 541-4 Saiho, Nakazato, Naka-Uonuma, Niigata 949-8400, JAPAN

Abstract: Snow pit studies have been carried out intermittently from January to April, 1995, at Saiho, Niigata, a temperate region in Japan, where the snow layer is mostly wet. Isotope analysis has been made of snow samples collected at every pit, and of precipitation and runoff water samples during the observation period. The mass of individual snow layers did not change with time, except at the bottom layer, but isotope ratio did change with time, increasing at shallow layers and decreasing at deeper layers. The $\delta^{18}O$ depth profile, however, was apparently preserved.

Introduction

Ice core analysis is one of the most powerful tools for reconstructing past climate [1]. Isotope records from ice cores have been considered to indicate past temperature variations. A recent finding [2] showing large and abrupt climate fluctuations revealed by analyzing the isotope profile along two ice cores from Greenland, is a good example. For reconstructing past climate from isotope records, however, the relationship between isotope content and temperature needs to be carefully evaluated [3]. Successful results have already emanated from the polar ice cores. In addition, ice core retrieval and analysis has been carried out lately at high elevation sites in mid- to low-latitude areas [4, 5]. The following two problems need to be examined, however, to obtain past temperature variations from isotope records in ice cores from these areas [6]. 1) The source area of water vapor, and its path to the precipitation site where the snow is formed and deposited need to be studied in detail. They must be taken into consideration in interpreting the isotope data and they are not as simple as in the polar areas. 2) Post-depositional change in isotope content could be significant and needs to be assessed, since glacier ice is frequently subjected to thaw/freeze cycles at many sites in low- to mid-latitude areas, and the presence of liquid water will accelerate the change of isotope content.

The first of these problems is solved by examining the distribution of isotope content in precipitation, water vapor and terrestrial water systematically over a wide area, and by examining the distribution pattern in combination with model calculations. This was planned and implemented within the framework of GAME (GEWEX in Asian Monsoon Experiment), a sub-program of

Physics of Ice Core Records
Edited by T. Hondoh
Hokkaido University Press, 2000, Sapporo

GEWEX (Global Energy and Water Cycle Experiment) [6]. For the second problem, the present study aims at examining the evolution of isotope content in a snow pack at the melting point.

Arnason et al. [7] and Krouse et al. [8] note that a $\delta^{18}O$ vertical profile does not change significantly with time in a snow pack at the dry snow zone or superimposed ice zone, but that it does tend to become uniform in a wet snow zone. Watanabe et al. [9] have also shown that the temporal change of precipitation $\delta^{18}O$ is preserved in a snow pack at sub-zero temperatures. Suzuki [10] examined the temporal change of both melt water and snow pack $\delta^{18}O$ at 0 °C. He found that the $\delta^{18}O$ value of the total snow pack as well as of the melt water increased as the melting season proceeded. He did not examine the change quantitatively, however, in relation to isotope balance of the snow pack.

Observations

Field observations were carried out from 10 January to 23 April 1995 on a flat area at altitude 445 m in Saiho, Nakazato-mura, Nakauonuma-gun, Niigata Prefecture, Japan. Snow pit studies were made intermittently, a total of 13 times during the observation period. The frequency of observations was 7 to 10 days. Snow thickness, air temperature, humidity, precipitation amount and global radiation were monitored continuously at the observation site for the full period. A lysimeter was also installed at the snow-ground interface and the amount of discharged water from the snow pack was measured, although it did not work successfully.

Figure 1 shows the variation of snow depth during the observation period. Vertical solid lines indicate the dates of the snow pit study. The figure also shows the distribution of snow temperatures at 0 °C (wet snow) roughly 0 to –0.1 °C, and below 0 °C (dry snow). It indicates that dry snow only was observed near the surface of the snow pack up till the end of February at Saiho, Niigata. In other words, the snow was roughly at 0 °C throughout the layers and throughout the observation period.

When deposited snow layers are kept at 0 °C for such a long time, snow particles tend to become granular, and the boundaries between neighboring snow layers become less identifiable. Markers to identify snow layer boundaries were hence placed at the surfaces on the 10 and 26 January, and on the 6, 20 and 27 February. These markers were then located at each snow pit study. The positions of the markers are shown with open circles in Fig. 1 in which curves connecting individual markers are also drawn. The following analysis is made of respective layers sandwiched by markers: layer A, B, C, D and E from the top downward. For example, layer A consists of snow deposited in the period between 20 February and 3 March.

The snow pit study includes stratigraphic observation, measurement of snow density, snow particle size, snow temperature and water content when wet at every 10 cm depth interval. Snow samples were collected during the pit study, also at 10 cm intervals. Melt water was occasionally sampled with the lysimeter. Precipitation samples were collected after 10 February. Prior to 10 April, precipitation was mostly as snow. On 13 February and 22 March it was sleet. These samples were subjected to isotopic analysis.

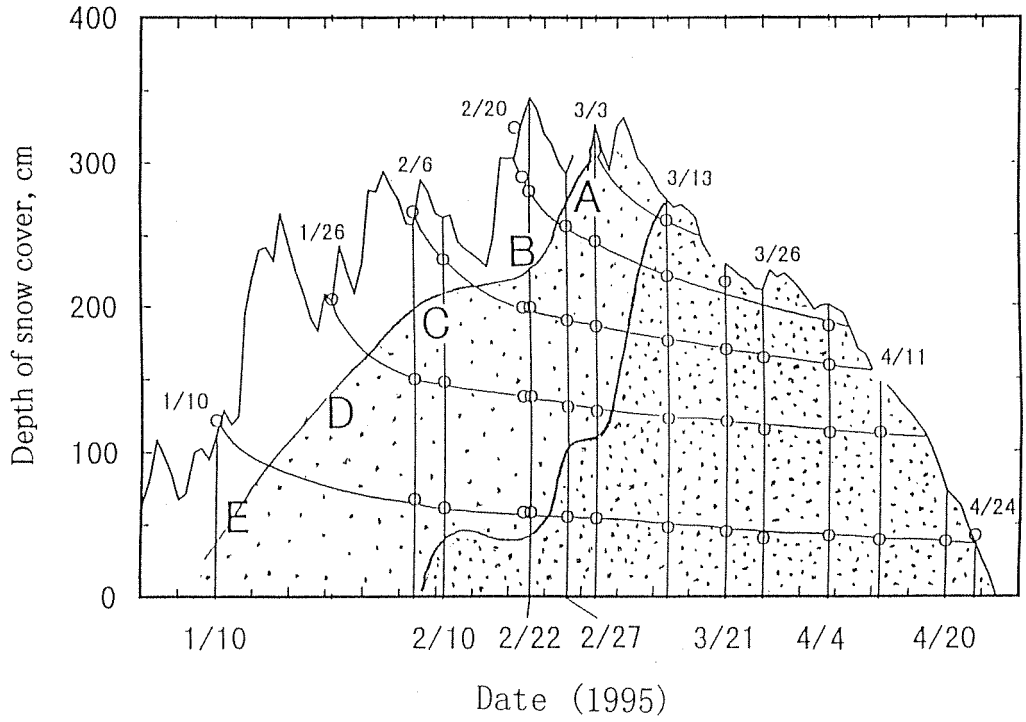

Figure 1: Depth of snow cover at Saiho, Niigata from January to April 1995. Vertical lines indicate the dates of snow pit studies. Open circles show the locations of markers buried in the snow between different layers, A, B, C, D and E. The area with a high density of dots is wet snow (temperature = 0 °C). The less densely shaded area is the area with temperatures in the range 0 to –0.1 °C. The un-shaded area is dry snow (temperature < 0 °C).

Water balance of the snow pack

Vertical water balance (one dimension) per unit area of the snow pack is expressed by

$$S_t - S_0 = \int_0^t P dt - \int_0^t R dt - \int_0^t V dt \qquad (1)$$

where S_0 and S_t are the total water mass of the snow pack including solid and liquid water at time 0 (when the observation was initiated) and t respectively. P, R and V are precipitation, discharge and evaporation per unit time respectively.

Figure 2 shows the water balance of the snow pack, S_t, and the cumulative precipitation. Because direct measurement by the lysimeter was not successful, and neglecting the contribution of evaporation, the temporal change of cumulative discharge was calculated from Equation 1. The result is also plotted in Fig. 2.

Figure 3 shows the temporal change of the water equivalent mass of individual layers. The mass is roughly constant for layers C and D, indicating that melt water at the surface has moved down through the layers underneath without any local freezing or melting which could cause a mass change. Also, water content did not change significantly with time in these layers, although it is not shown here. The mass for

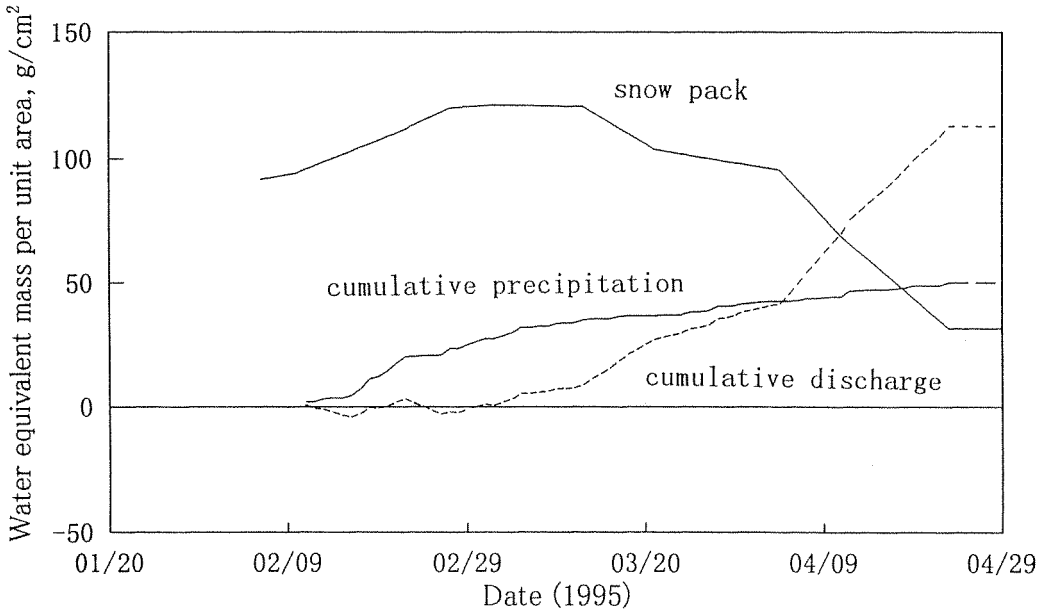

Figure 2: Water balance of the snow pack. The mass per unit area of the snow pack obtained from the snow depth and the profile of snow density. Cumulative discharge was calculated from Equation 1, neglecting the mass loss by evaporation.

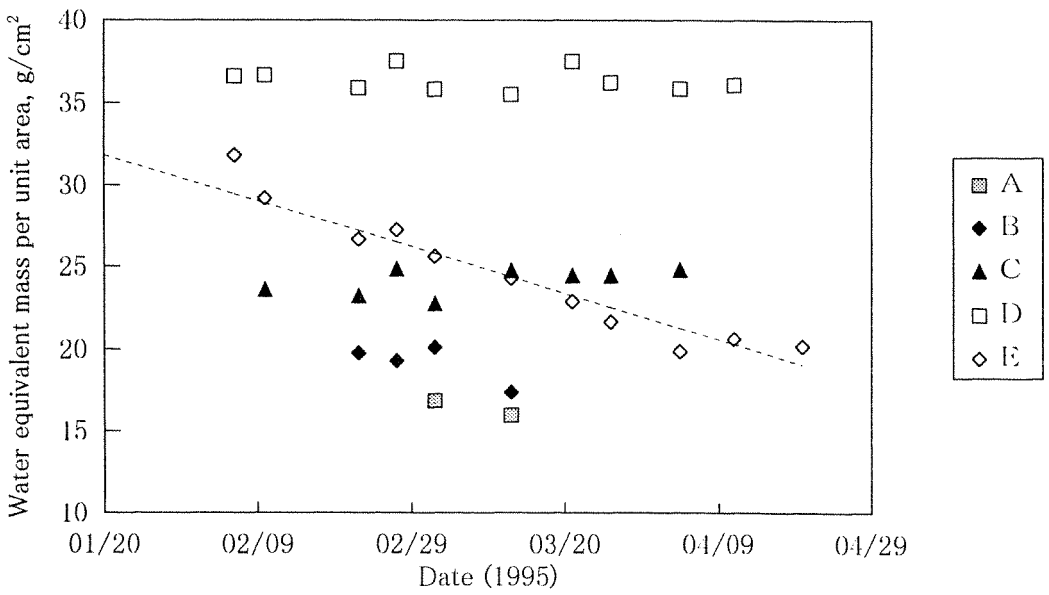

Figure 3: Temporal change of water mass of individual snow layers. Mass is approximately conserved for layers C and D, although a slight decreasing trend can be seen for layers A (two data points only) and B. A significant decrease (shown by the linear regression) is found for layer E as a result of bottom melting.

layers A (two data points only) and B shows a slight decreasing trend. The water content in these near-surface layers could have decreased slightly with time, because the growth rate of snow grains would be rather large soon after deposition, leading to a decrease in water holding capacity. The water content data also showed a decreasing trend, although the scatter of the data was large. The bottom layer, E decreased in mass significantly, as a result of bottom melting. Linear regression indicates a melt rate of 1.4 mm/day, which is equivalent to a heat flux of about 5 W/m^2, based on the water content data of 3 to 5 % in the bottom layer. This value for the geothermal heat flux is 2 to 3 times higher than the values obtained by Kojima [11, 12] at a site in Hokkaido. It is compatible with the value reported by Kojima and Motoyama [13] since the geothermal heat is considered larger in snowy regions in Honshu Island, where Saiho, Niigata is located, than in Hokkaido Island [14].

δ^{18}O of the snow pack

The oxygen isotope ratio, δ^{18}O for samples taken from the snow pack is contoured within the snow depth plot in Fig. 4. δ^{18}O values range from −8 to −14 ‰. Horizontal stratification seen in the figure indicates that the vertical profile is largely preserved even in the wet snow layers. The largest δ^{18}O peaks in the vertical profile decrease gradually with time, corresponding with the settling of snow layers. Figure 5 shows the temporal change in δ^{18}O (weighted-average) for individual layers. Although the data are scattered, we can see

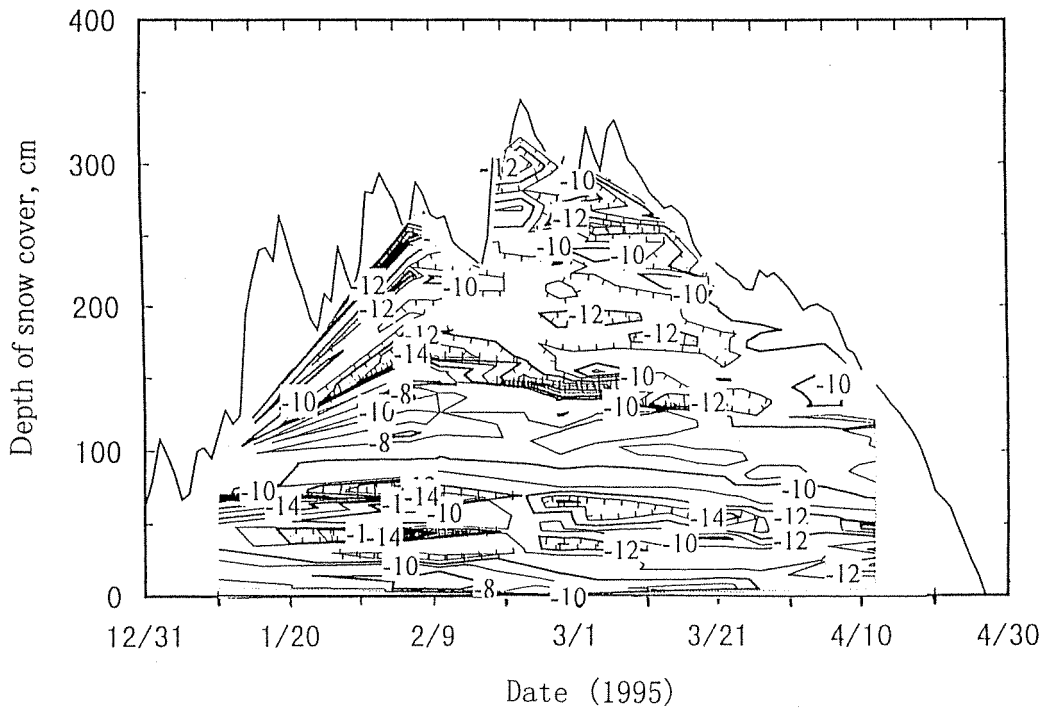

Figure 4: The δ^{18}O value of the snow is contoured within the plot of snow cover depth as a function of time. Horizontal stratification indicates that isotope value is approximately preserved.

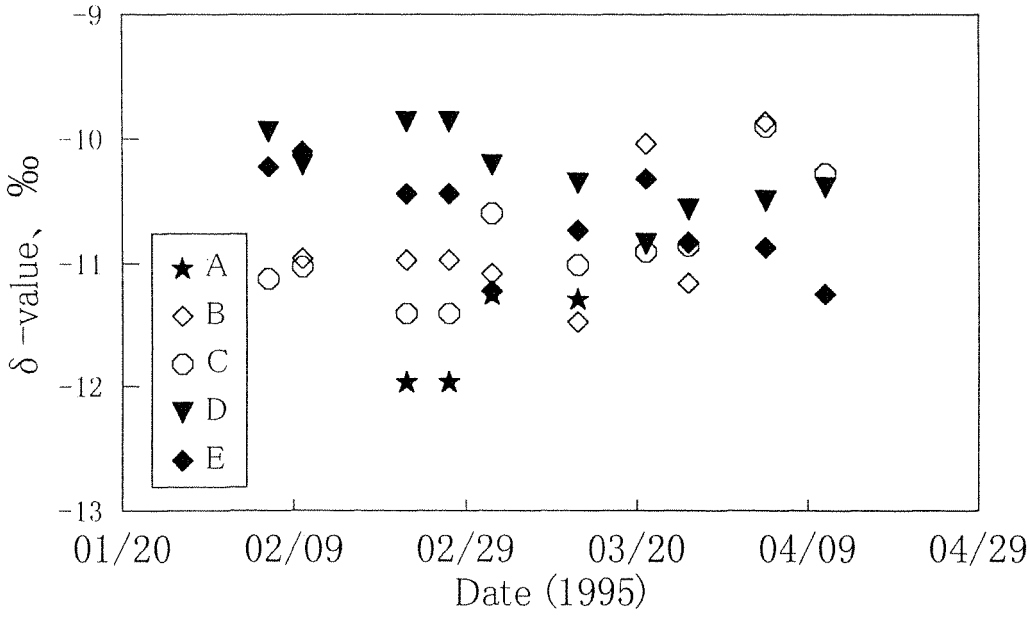

Figure 5: Temporal change of $\delta^{18}O$ for each snow layer. $\delta^{18}O$ increases with time in shallow layers (A, B and C) but decreases in deeper layers (D and E).

that $\delta^{18}O$ increases with time in the shallower layers A, B and C, but decreases in the deeper layers D and E. This decrease is not considered to reflect a change in snowfall $\delta^{18}O$ because the data shown in Fig. 5 indicate temporal changes for the snow after deposition.

When snow particles are in contact with liquid water, isotopic fractionation takes place at the interface. Heavy isotopes tend to concentrate on ice and liquid water is depleted. This has been confirmed by experimental work by Nakawo et al. [15]. As snow melts at the surface of the snow pack, melt water becomes depleted of heavy isotopes, and in lower layers snow particles are enriched. This results in a gradual increase in $\delta^{18}O$ in snow layers near the surface. As melt water percolates downward, the depletion proceeds further, and the isotopic exchange takes place between snow particles and the very depleted water, so that isotope content decreases with time at lower layers.

Figure 6 is the $\delta^{18}O$ – δD diagram for the individual layers A, C and E. The gradient, $\delta D/\delta^{18}O$ decreases with depth: 10.6 for A, 8.8 for B, 7.5 for C, 7.8 for D and 6.9 for E. Nakawo et al. [15] derived a relation to express the gradient $\delta D/\delta^{18}O$ in terms of the initial deuterium and ^{18}O content of both solid particles and liquid water. Their expression is approximated to give the gradient as

$$\frac{\delta D}{\delta^{18}O} \approx \frac{\left((\delta D_{i0}+1)-\alpha_D(\delta D_{w0}+1)\right)}{\left((\delta^{18}O_{i0}+1)-\alpha_{18O}(\delta^{18}O_{w0}+1)\right)} \qquad (2)$$

where δD_{i0} and $\delta^{18}O_{i0}$ are the initial deuterium and ^{18}O content for ice portion, and δD_{w0}, $\delta^{18}O_{w0}$ are the initial deuterium and ^{18}O content for water, and α_D and α_{18O} are the fractionation factor for deuterium and ^{18}O respectively. They applied the equation to a

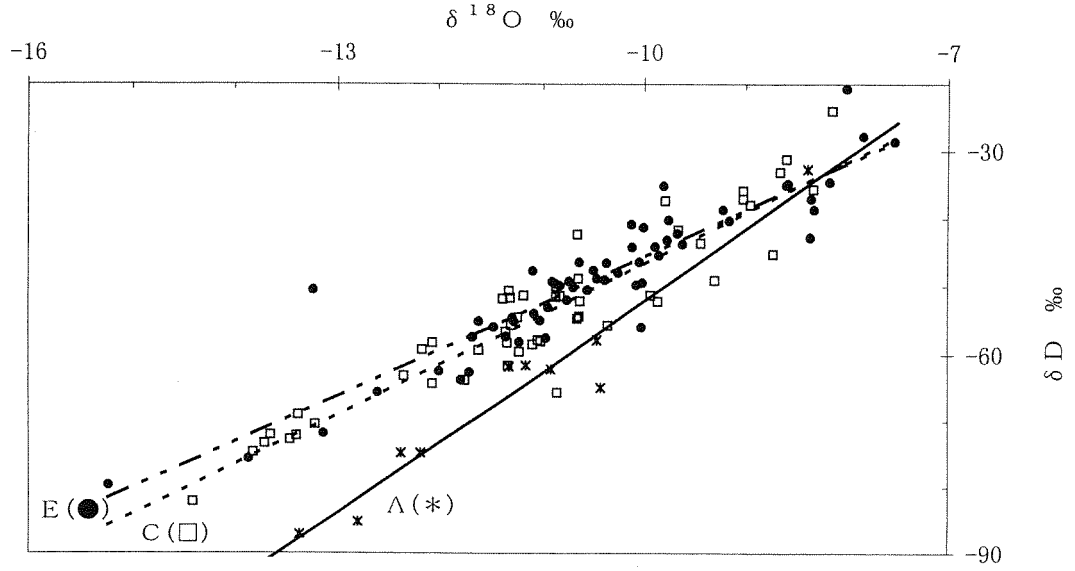

Figure 6: $\delta^{18}O$ – δD diagram for the change of isotope content in individual layers, A, C and E. The gradient of the regression lines, $\delta D/\delta^{18}O$ decreases with increasing depth.

snow-water system in which snow particles were completely soaked in water. Their relation was derived for a mass mixing ratio of 0.5. The ratio of water in our wet snow pack is much smaller than 0.5. Nonetheless, we apply the relation assuming the initial isotope ratios for both water and snow are equal to the average value for the entire snow pack at the beginning of the observation period. This corresponds to the interaction between snow particles and water at layers near the surface where melting takes place. The calculation results in a $\delta D/\delta^{18}O$ value of about 7. As melt water percolates downward, the $\delta^{18}O$ of the water decreases as a result of isotope exchange between solid and liquid water. If the isotope content is the same as initially for snow particles but with depleted water, then the gradient reduces (It changed to 6, 5 and so on with decreasing isotope content in the water). Although the above calculation does not agree with the observed gradient quantitatively, the decreasing trend in the gradient with depth can be explained qualitatively, since the depletion of water is accelerated with depth.

Isotope balance of the snow pack

Figure 7 shows the daily precipitation and $\delta^{18}O$ value. Precipitation is relatively large in February, and decreases toward the end of the season. The $\delta^{18}O$ value varied in the range 0 to −20 ‰. Figure 8 shows the variation of $\delta^{18}O$ measurements on discharge water collected intermittently. The values are stable in February while the discharge amount is small (Fig. 2). The $\delta^{18}O$ value in this period is −8 to −9 ‰, i.e. similar to values for snow at the bottom of the snow pack (Fig. 4). This indicates that discharge water in this period is due to bottom melting of the snow pack. In March, while melting is pronounced, $\delta^{18}O$ varied in the range −5 to −11 ‰. Over the total period a small

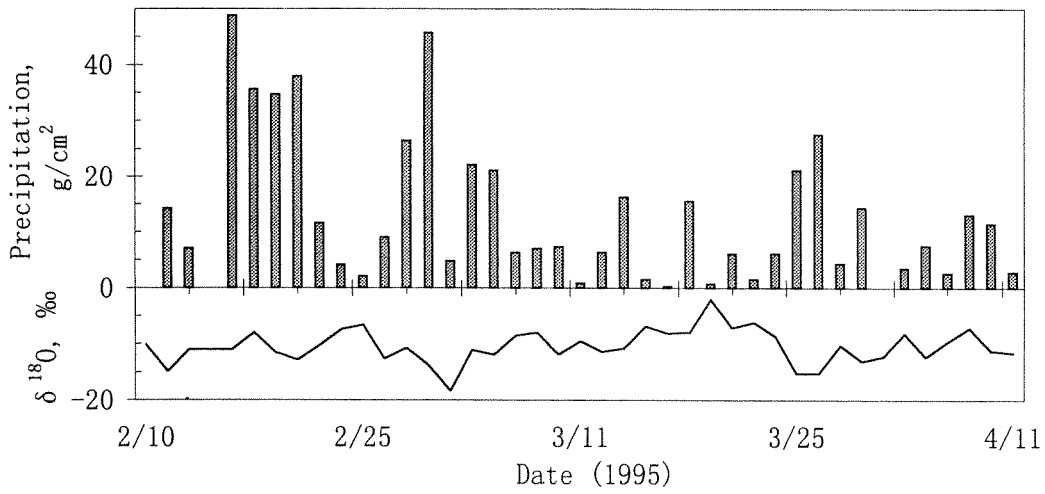

Figure 7: The variation of daily precipitation and precipitation $\delta^{18}O$ value during the observation period.

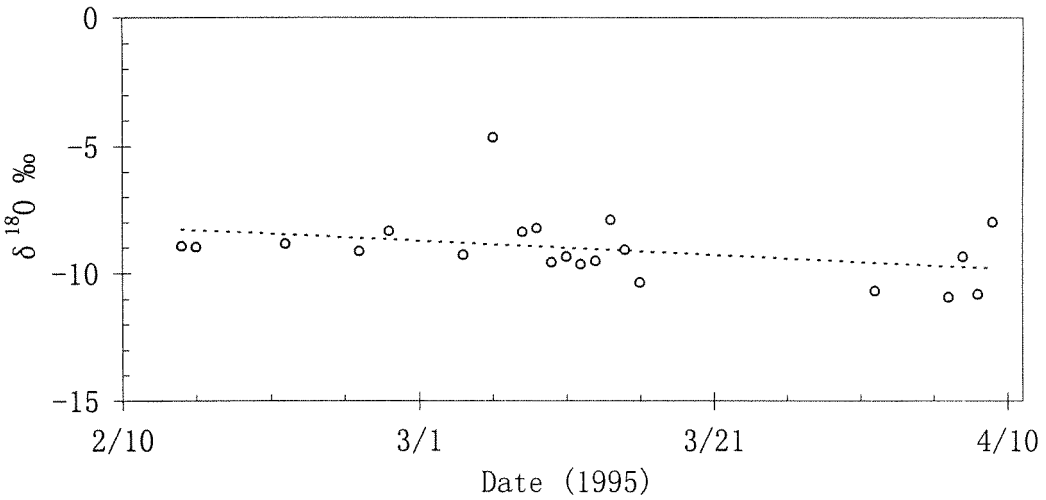

Figure 8: The variation of discharge water $\delta^{18}O$ value, showing a gradual decreasing trend. $\delta^{18}O$ values of –8 to –9 ‰ found before the melt period (within February) correspond with the $\delta^{18}O$ values near the bottom of the snow pack (see Fig. 4).

decreasing trend can be seen as indicated by the linear regression in Fig. 8.

The total oxygen isotope amount in the snow pack at time t can be expressed by

$$I_t = \sum_{A}^{E} \delta M \tag{3}$$

where δ and M are $\delta^{18}O$ value and water mass respectively for each of the layers A, B,

C, D and E. The change in I_t as a function of time is shown by triangles in Fig. 9. Isotopic balance can be expressed (cf. Equation 1) by

$$I_t - I_0 = \int_0^t \delta_P P dt - \int_0^t \delta_r R dt - \int_0^t \delta_v V dt \qquad (4)$$

where I_0 is initial isotope amount in the snow pack at time 0. δ_P, δ_r and δ_v are the $\delta^{18}O$ values for the precipitation, discharge water and evaporated vapor respectively, and V is the rate of evaporation from the snow pack. Neglecting the third term of the right hand side, I_t is calculated from Equation 4. P, R, δ_P and δ_r are obtained from Figs. 7 and 8, δ_r by interpolation from the linear regression of Fig. 8. The result of this calculation is shown by the solid curve of Fig. 9. The calculation roughly agrees with the data derived from the snow pit study early in the season. Later in the season, however, Equation 4 gives a smaller total isotope amount than indicated by the snow pit data.

The discrepancy could be due to neglecting the evaporation effect. When evaporation takes place from the snow surface, δ_v is given by

$$\delta_v = \frac{1}{\alpha}(1+\delta_s) - 1 \qquad (5)$$

where δ_s is the $\delta^{18}O$ value for the surface snow, and α is the fractionation factor. δ_v is about −22 ‰ for α = 1.011, an equilibrium factor provided by Dansgaard [16] and δ_s = −11 ‰ (a typical value at the surface of the snow pack). In general, however, kinetic effects need to be considered when examining fractionation in association with evaporation,

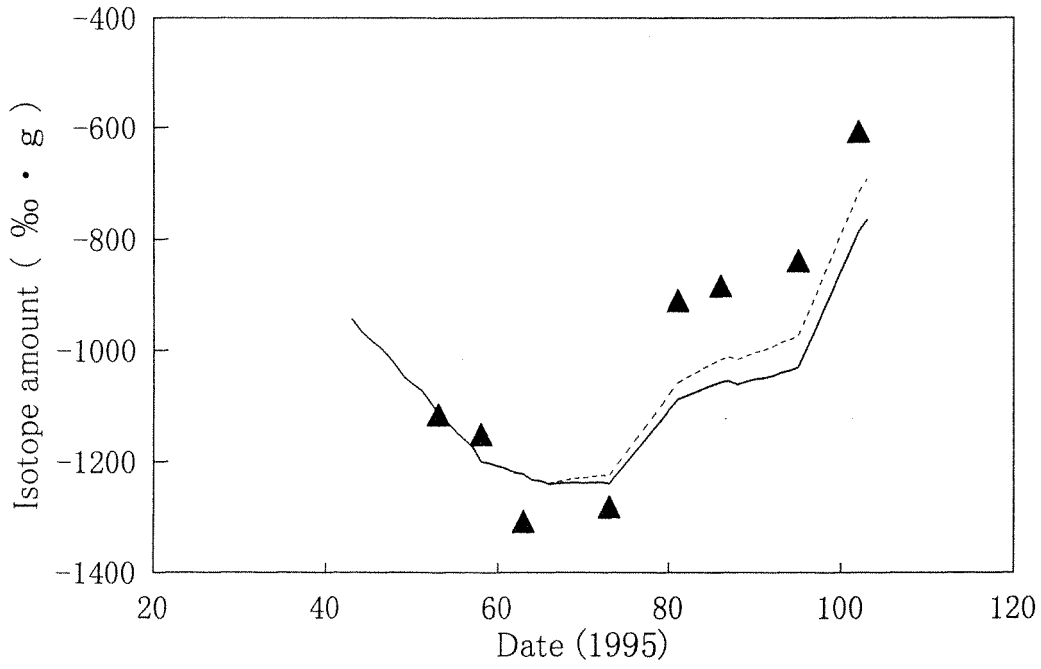

Figure 9: The variation of total isotope amount in the snow pack. Solid triangles indicate the observed results obtained for the $\delta^{18}O$ values for the snow pit samples. The solid curve is derived from Equation 4 with no evaporation assumed. The dashed curve does take evaporation effects into account.

and a larger value for the fractionation factor is expected for the equilibrium state. Nakawo and Takahara [17] for example, obtained a value of 1.03 for the fractionation factor with kinetic effects in an arid area of western China. At Saiho where the observations for the present study were carried out, the humidity is higher than in western China, and the fractionation factor is expected to be smaller. A value of 1.02 is therefore used here for a rough estimate of the contribution of evaporation to isotope balance, since no reliable value is available in our observations. This value results in a δ_v value of −30 ‰. The rate of evaporation is considered to be about 0.5 mm/day (and a maximum of 1 mm/day) in the observation area during the season. With these values, and taking into account the effect of evaporation, the total isotope amount in the snow pack was re-calculated, again using Equation 4. The result is closer to the observations as shown by the dashed curve of Fig. 9. The calculation is approximate only, since no reliable fractionation factor or evaporation rate are available. It does indicate however, that the effect of evaporation should not be neglected in examining isotope balance, although it is often neglected in examining water balance.

Concluding remarks

The vertical profile of $\delta^{18}O$ value is roughly preserved for about 2 months, even in the wet snow pack (Fig. 4). This is encouraging for the analysis of ice cores from low- to mid-latitude glaciers. However $\delta^{18}O$ is subjected to post depositional changes through interactions between snow particles and liquid water. It is crucial, therefore, to understand the fractionation between the two phases quantitatively. Evaporation (or condensation) could play a role in post depositional changes of isotope content. The formation of ice layers due to re-freezing of percolated water also needs to be examined in order to establish a reliable relation between climate and isotope record in ice cores, although ice layer formation did not take place in our wet snow pack.

Acknowledgements

The authors wish to thank the following colleagues for supporting the field observations: K. Seko, K. Fujita, T. Kadota, B. Rana, L. E. Goodrich, A. Sakai and H. Yabuki of Nagoya University, M. Shimizu and K. Goto-Azuma of National Institute for Earth Science and Disaster Prevention, T. Aoki of Meteorological Research Institute, and J. Kubota of Tokyo University of Agriculture and Technology. Thanks are due to N. Yoshida and O. Abe of Nagoya University, who assisted with the measurements of deuterium content of the samples. The authors are grateful also with anonymous reviewers, whose comments improved the paper significantly.

References

1. H. Oeschger and C.C. Langway Jr., *The environmental Record in Glaciers and Ice Sheets*, John Wiley & Sons (1989).
2. W.S. Dansgaard, S.J. Johnsen, H.B. Clausen, D. Dahl-Jensen, N.S. Gundestrup, C.U. Hammer, C.S. Hvidberg, J.P. Steffensen, A.E. Sveinbjornsdottir, J. Jouzel and G. Bond, *Nature*, 364, 218-220 (1993).

3. G. de Q. Robin, *The climatic record in polar ice sheets*, Cambridge University Press (1983).
4. L.G. Thompson, E. Mosley-Thompson, J.F. Bolzan and B.R. Koci, *Science*, 229, 971-973 (1985).
5. L.G. Thompson, X. Wu, E. Mosley-Thompson and Z. Xie, *Ann. Glaciol.*, 10, 1-5 (1988).
6. M. Nakawo, *Proceedings of the International Workshop on the Climate System of Monsoon Asia*, 228-229 (1996).
7. B. Arnason, Th. Buason, J. Martinec and P. Theodorsson, *IAHS Publication*, 106, 299-312 (1973).
8. H.R. Krouse, H. Hislop, H.M. Brown, K. West and J.L. Smith, *IAHS Publication*, 118, 242-247 (1977).
9. O. Watanabe, N. Kanamori, A. Sugimoto, H. Iida and K. Higuchi, *J. Earth Sci. Nagoya Univ.* 34, 67-108 (1986).
10. K. Suzuki, *Seppyo*, 55, 335-342 (1993).
11. K. Kojima, *Low Temperature Science, Ser. A*, 39, 101-108 (1980).
12. K. Kojima, *Low Temperature Science, Ser. A*, 41, 99-107 (1982).
13. K. Kojima and H. Motoyama, *Ann. Glaciol.*, 6, 276-277 (1985).
14. H. Motoyama, *J. App. Meteorol.*, 29, 1104-1110 (1990).
15. M. Nakawo, S. Chiba, H. Satake and S. Kinouchi, *Ann. Glaciol.*, 18, 129-134 (1993).
16. W. Dansgaard, *Tellus*, 16, 436-467 (1964).
17. M. Nakawo and H. Takahara, *Bull. Glacier Res.*, 6, 9-13 (1988).