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Scientific note

Stable oxygen isotope ratio observed in the precipitation at Ny-Ålesund, Svalbard

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Abstract: The stable oxygen isotope ratio ($\delta^{18}O$) in ice cores and surface snow pits samples obtained in Svalbard does not show a seasonal variation clearly. In preceding research, it has been reported that $\delta^{18}O$ is change by melting of the snow layer. However, to what degree $\delta^{18}O$ in snow changes after accumulation is not clear. We carried out the observation in Ny-Ålesund, northwestern part of Svalbard, in the 1993/94 winter season and 1994 summer season in order to investigate the initial values of $\delta^{18}O$ in precipitations.

Values of $\delta^{18}O$ both in winter and summer were almost same (−4∼−24‰) as well as the average of each season was also almost same (−12.9‰, −12.3‰), which shows that $\delta^{18}O$ in precipitations does not have a clear seasonal variation.

It is known that the value of $\delta^{18}O$ in precipitation changes by the condensation temperature of water vapor. Temperature, at the height where the wettest air mass existed over northwestern Svalbard, changed almost same range in winter and summer, when the rain fell at Ny-Ålesund. And, there was a better correlation between $\delta^{18}O$ in precipitation and temperature of moist air masses than $\delta^{18}O$ in precipitation and air temperature at the observation site both in winter and in summer.

Therefore, fluctuation of $\delta^{18}O$ seemed to show the almost equal, even if the season was different. This fact can be regarded that one of the reasons why $\delta^{18}O$ of ice core drilled in Svalbard does not show the seasonal variation clearly.

1. Introduction

Dansgaard (1953) has indicated that past temperature is estimable by analysis of the stable oxygen isotope ratio ($\delta^{18}O$) in ice cores. The values of $\delta^{18}O$ in ice cores drilled on glaciers have been used for discussing paleoclimate. For example, climatic changes for several hundred thousand years have been clarified from ice cores obtained in Antractica and Greenland (*e.g.* Petit *et al.*, 1999; Dansgaard *et al.*, 1993). And seasonal variation of $\delta^{18}O$ was preserved for the last 8300 years in an ice core from Camp Century, northwestern Greenland (Johnsen *et al.*, 1972).

In Svalbard, research on the reconstruction of the climatic variation from $\delta^{18}O$ analysis has also been carried out. Goto-Azuma *et al.* (1995) drilled on Snøfjellafonna, northwestern Spitsbergen until 84 m depth, and found that the seasonal variation of $\delta^{18}O$ had partially been lost. This confirmed the finding that the seasonal variation of $\delta^{18}O$ was not recognizable in analysis of an ice core from Vestfonna on Nordaustlandet by Punning *et al.* (1986). The seasonal variation of $\delta^{18}O$ did not show clearly in analysis of surface snow pit samples obtained in Austfonna on Nordaustlandet as well as ice cores (Iizuka *et al.*, 2000). The reason why the seasonal variation did not appear clearly in $\delta^{18}O$ was reported by Goto-Azuma *et al.* (1995) and Punning *et al.* (1986), as the snow melting after the accumulation on the glacier. However there is few data on the change of $\delta^{18}O$ in precipitation in these regions. We collected the precipitation in Svalbard in different season for the basic discussion.

2. Samples and methods

Precipitation sampling on the glacier over several months is not so easy. Therefore, precipitation samples were collected at Ny-Ålesund (78°56′N, 11°52′E, 35 m a.s.l.) in northwestern Svalbard (Fig. 1), where there is an observation base of the National Institute of Polar Research, and it is possible to stay for a long decade and to freeze the precipitation samples for the preservation on chemical analysis. Meteorological observations are carried out there regularly, and can be easily compared with results of analysis of the collected precipitation.

The investigation point at Ny-Ålesund is shown in Fig. 2. It is 1.6 km from the center of town and 0.2 km from the coast of Kongsfjord.

The precipitation sampling was carried out in winter 1993/94 and summer 1994. It is shown in Table 1. Precipitation could not be collected from December 16, 1993 to Januruary 31, 1994, because there was no investigator at Ny-Ålesund. During the obsevation period, there was precipitation for 36 days in winter and for 21 days in summer. Precipitation was observed as snowfall and rain both in winter and summer.

The sampling procedures between for snowfall and for rain were different. In case of snowfall, we collected fresh snow in a pre-cleaned plastic bag. Rain was concentrated in a

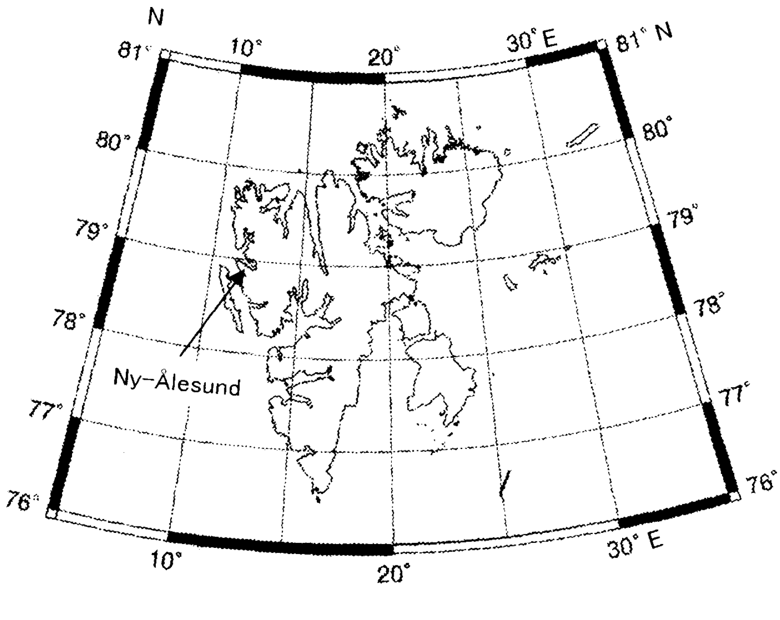

Fig. 1. Location of study area.

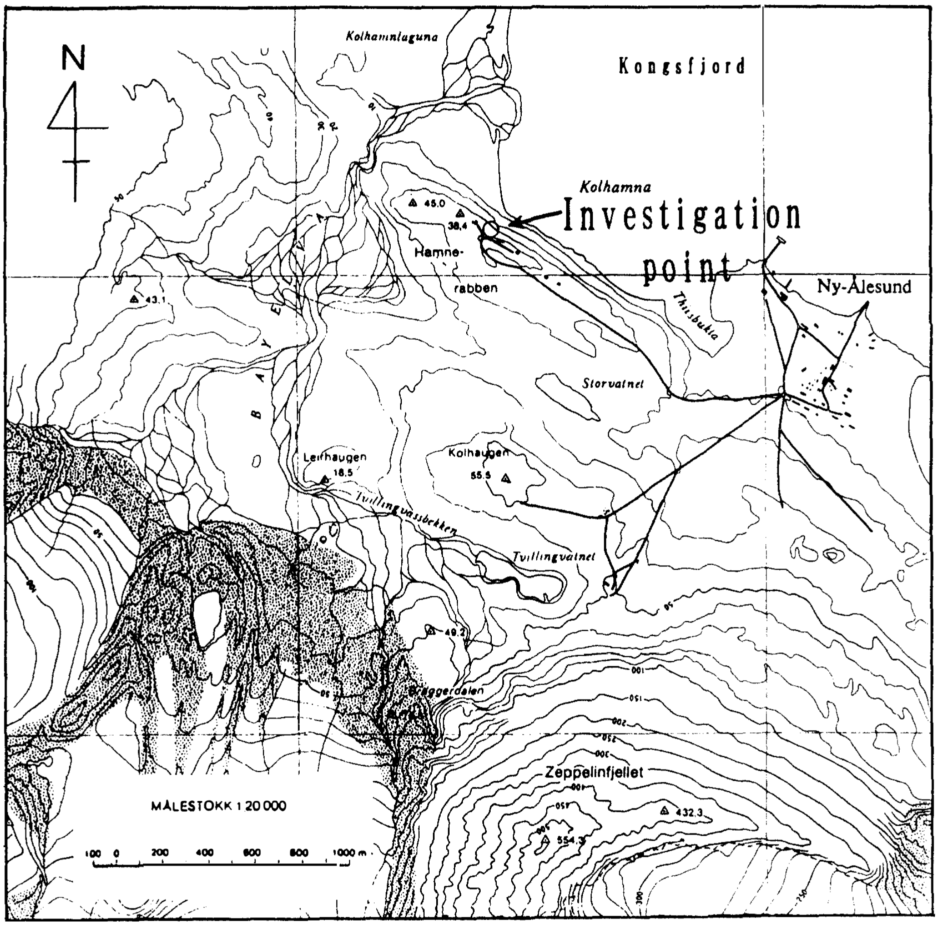

Fig. 2. Landforms and investigation point at Ny-Ålesund. The region where the mesh splashed is the end moraine.

Table 1. Outline of the observations.

	Date	The number of days of observation	The days with precipitation	The days of sample obtained
Winter 1993/94	1993.12.3~12.15 1994.2.1~3.15	56	36	31
Summer 1994	8.23~9.18	27	21	15

pre-cleaned plastic bottle with a funnel of diameter 0.3 m. We collected 100 m*l* samples to have enough for not only measurement of $\delta^{18}O$ but also determination of the chemical composition. And in the case of snowfall and rain, we collected samples every 6 hours at the time of meteorological observations. It is desirable to collect each sample in as short a time as possible because in Ny-Ålesund, the climate dramatically changes in a day during the polar night (Hanssen-Bauer *et al.*, 1990). During the summer, precipitation of 0.1 mm for 6 hours was necessary in order to collect a 100 m*l* sample. During the winter, it was not possible to collect a sufficient sample of 0.1 mm precipitation for 6 hours, as a sample was collected when there was 0.1 mm precipitation for 12 hours. There was sufficient precipitation for sampling during 31 days in winter and 15 days in summer.

Each of the snowfall samples was melted in the laboratory at Ny-Ålesund and was kept in a pre-cleaned plastic bottle. After that, samples were kept frozen until measurement of $\delta^{18}O$. $\delta^{18}O$ was measured by mass spectrometer (Delta-E, finnigan-mat). The analytical error was ±6%.

3. Characteristics of climate in this study period

Meteorological data were measured by the Norwegian Meteorological Institute 3 times per day (0700, 1300 and 1900; GMT+1) at Ny-Ålesund (Fig. 2). Sea-level pressure, surface air temperature, relative humidity and precipitation are shown in Table 2. Precipitation was observed 2 times per day (0700 and 1900; GMT+1). The deviation from the average of each quantity in each period from 1975 to 1989 is indicated in parentheses in Table 2, as well as the standard deviation of temperature within each period.

The deviation from surface mean temperature between 1975 and 1989 (−2.3) exceeded the standard deviation (0.6) only in August. The deviation of surface mean temperature in other months fell below the standard deviation (Table 2). Especially, the mean temperature during the winter of 1993/94 (December 3, 1993–March 15, 1994) agreed with the mean temperature between 1975 and 1989. And the deviation throughout the summer of 1994 (−0.9: August 23–September 19) was smaller than the standard deviation from August to September (1.1). This means that there was hardly a difference of mean temperature between the whole observation period in the summer and from 1975 to 1989. Large negative deviation appeared in August. It was about 4 times of the standard deviation (Std). This is because mean temperature (1.6℃) for the final 9 days of observation in August was 1.1℃ lower than the mean temperature (2.7℃) during August.

The deviation from mean value of sea-level pressure in March in the first half of the study period was the greatest difference from mean values.

Precipitation in January and February showed the large absolute value of deviations from the mean value. In January, the precipitation became only 4% of the mean value. As

Table 2. Meteorological data. Deviations from mean values between 1975 and 1989 are shown in parentheses.

	Sea-level pressure (hPa)		Temperature (℃) mean		Std.*	min.	max.	Relative humidity (%)		Precipitation (mm)	
Winter 1993/94											
12/3–3/15	1000.8	(−0.8)	−13.8	(0.0)	4.0	−32.8	4.4	69.9	(−6.1)	144.6	(26.6)
12/3–31	1005.5	(−0.6)	−11.9	(1.0)	4.6	−25.1	−0.7	62.6	(−10.4)	26.9	(1.6)
1/1–31	1007.6	(1.4)	−15.0	(−0.6)	4.0	−26.0	−1.6	58.5	(−15.5)	1.1	(−26.9)
2/1–28	1013.9	(5.6)	−14.7	(−0.1)	3.8	−32.2	4.4	82.1	(3.1)	97.5	(56.5)
3/1–15	994.3	(−15.4)	−13.6	(−0.1)	3.7	−32.8	0.6	85.4	(7.4)	19.1	(−4.6)
Summer 1994											
8/23–9/19	1007.2	(−2.4)	0.8	(−0.9)	1.1	−3.9	7.7	81.2	(−2.3)	46.5	(9.3)
8/23–8/31	1002.5	(−9.1)	1.6	(−2.3)	0.6	−3.9	7.7	74.7	(−10.3)	17.8	(5.9)
9/1–19	1009.5	(1.8)	0.4	(0.9)	1.6	−3.3	4.4	84.2	(2.2)	28.7	(3.4)

*The standard deviation of mean temperature in each period.

the result, the relative humidity was 15.5% lower than mean value. On the contrary, the amount of precipitation in February became more twice larger than mean value. However, the relative humidity was only 3% larger than mean value.

4. Results and discussion

4.1. Relationship between $\delta^{18}O$ and surface air temperature

$\delta^{18}O$ of the precipitation in winter 1993/94 is shown in Fig. 3a, and that in summer 1994 is shown in Fig. 4a. Surface air temperature measured at Ny-Ålesund is presented in the same figures in order to compare with the fluctuation of $\delta^{18}O$.

In winter, $\delta^{18}O$ varied between −7.3‰ and −20.4‰ (Fig. 3a), and between −4.7‰ and −23.6‰ in summer (Fig. 4a). The fluctuating range of $\delta^{18}O$ was about 6‰ larger in summer than in winter. The averages of $\delta^{18}O$ were −12.9‰ in winter and −12.3‰ in summer respectively.

Dansgaard *et al.* (1973) illustrated the relationship between $\delta^{18}O$ in precipitation collected in the Arctic, and surface mean temperature. They reported that the annual mean of surface temperature and that of $\delta^{18}O$ appeared to be proportional. The $\delta^{18}O$ mean value

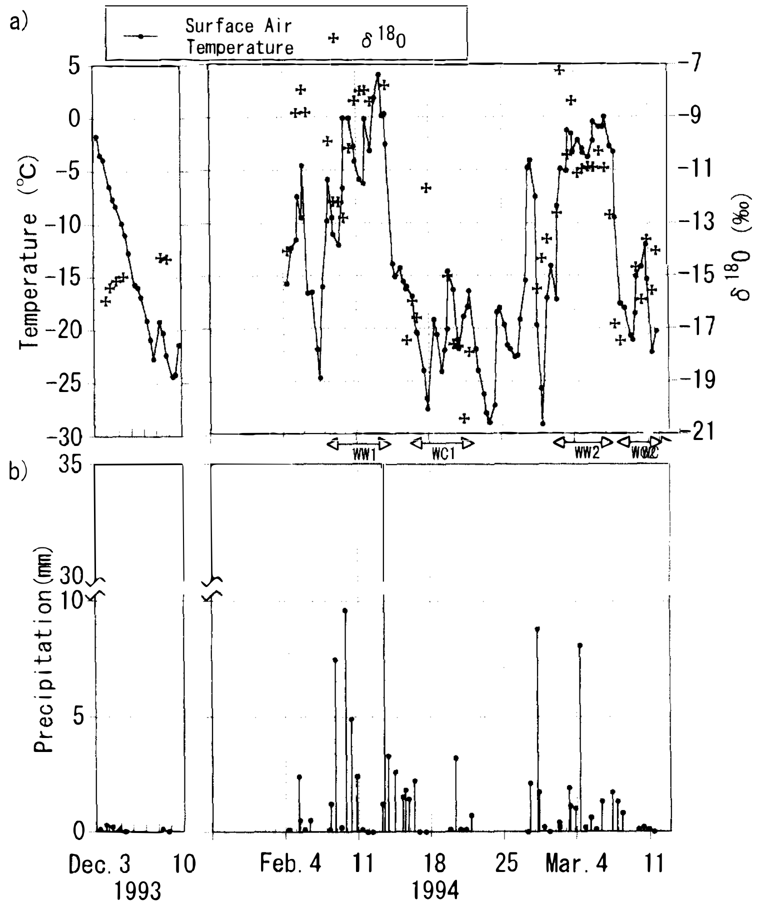

Fig. 3. $\delta^{18}O$ in precipitation, surface air temperature and precipitation at Ny-Ålesund in winter 1993/94. a) Surface air temperature and $\delta^{18}O$ in precipitation. b) Precipitation.

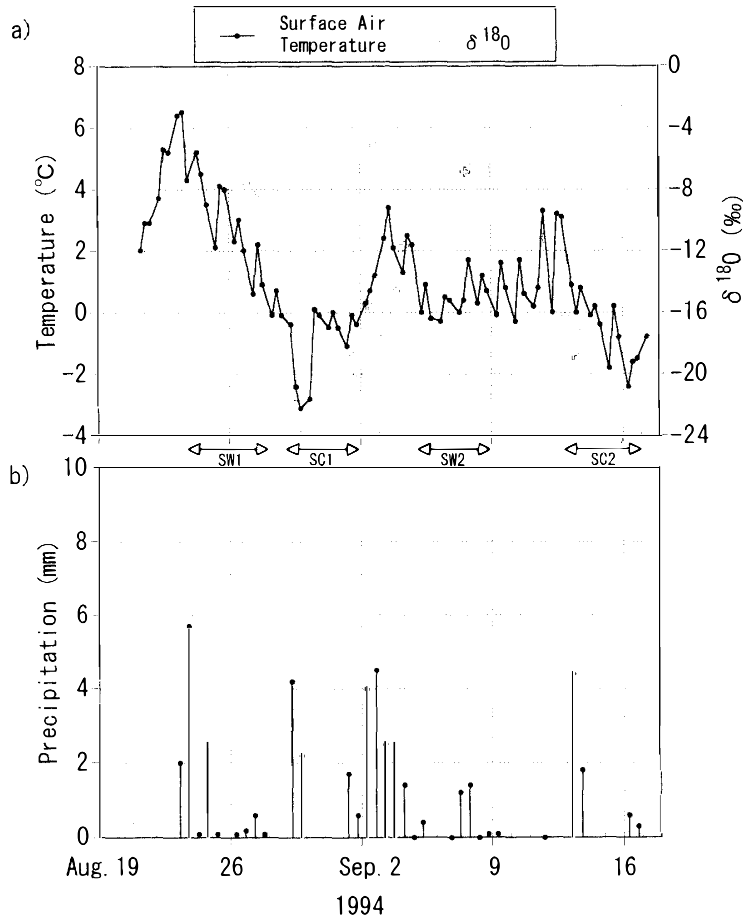

Fig. 4. *$\delta^{18}O$ in precipitation, surface air temperature and precipitation at Ny-Ålesund in summer 1994. a) Surface air temperature and $\delta^{18}O$ in precipitation. b) Precipitation.*

of all precipitation which we collected in Ny-Ålesund was −12.6‰, and surface mean temperature during the observation period was −8.9℃. This result agreed on proportional relationship of the other Arctic. Therefore, there appears to be correlation between the average of δ^{18}O in precipitation and surface mean temperature at Ny-Ålesund. We considered that the mean value of δ^{18}O in precipitation collected during this observation was almost usual at Ny-Ålesund because surface mean temperature was not so different from mean value between 1975 and 1989.

Fluctuation of δ^{18}O is related to temperature (Dansgaard, 1953). This means that features of the temperature during the observation period should reflect similar changes in δ^{18}O. Frequency distributions of δ^{18}O in Ny-Ålesund and surface air temperature in winter and summer are shown in Fig. 5. The distribution of δ^{18}O (Fig. 5a) in the summer was longer than that in winter and the difference of both average were only 0.6‰, as it was mentioned earlier. However, the temperature distribution in summer was smaller than the one in winter. Temperature during each precipitation fluctuated between −28.9℃ and 0.1℃ in winter and between −3.9℃ to 7.7℃ in summer (Fig. 5b). The variation was about three times larger in winter than in summer. Average temperature were −13.8℃ in winter

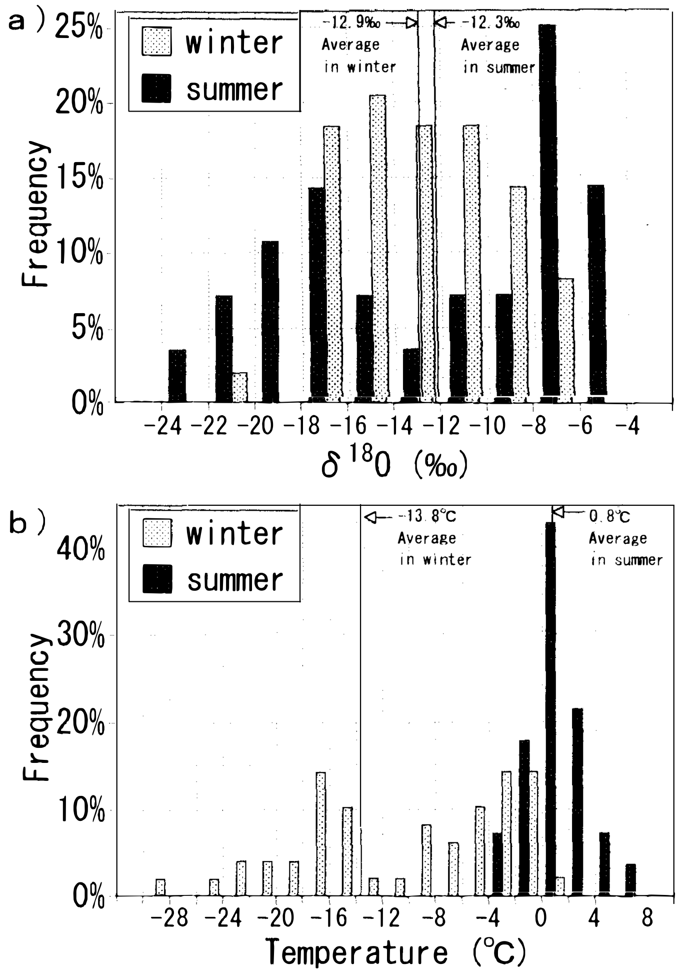

Fig. 5. Frequency distributions of $\delta^{18}O$ in Ny-Ålesund and surface air temperature in winter and summer. The total number of data are 49 in winter and 28 in summer. a) $\delta^{18}O$. b) Surface air temperature.

and 0.8℃ in summer, respectively. Large distribution of δ^{18}O occurs under the small distribution of temperature, and smaller distribution of δ^{18}O in winter under the large distribution of temperature, which indicated that the fluctuation of δ^{18}O does not depend upon only in the change of the surface air temperature.

4.2. Relationship between $\delta^{18}O$ and aerology data

In the previous section, we represented that it was difficult to explain the fluctuation of δ^{18}O only in the change of the local surface air temperature. Then, there is some possibility that the distribution of δ^{18}O depends under the atmospheric condition under the larger scale.

We used aerology data edited at NOAA-CIRES Climate Diagnostics Center. These data are compiled on a 2.5-degree latitude×2.5-degree longitude global grid and in 17 pressure levels (hPa: 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, 10). The grid point which is nearest to Ny-Ålesund is 80.0°N, 12.5°E. Temperature and

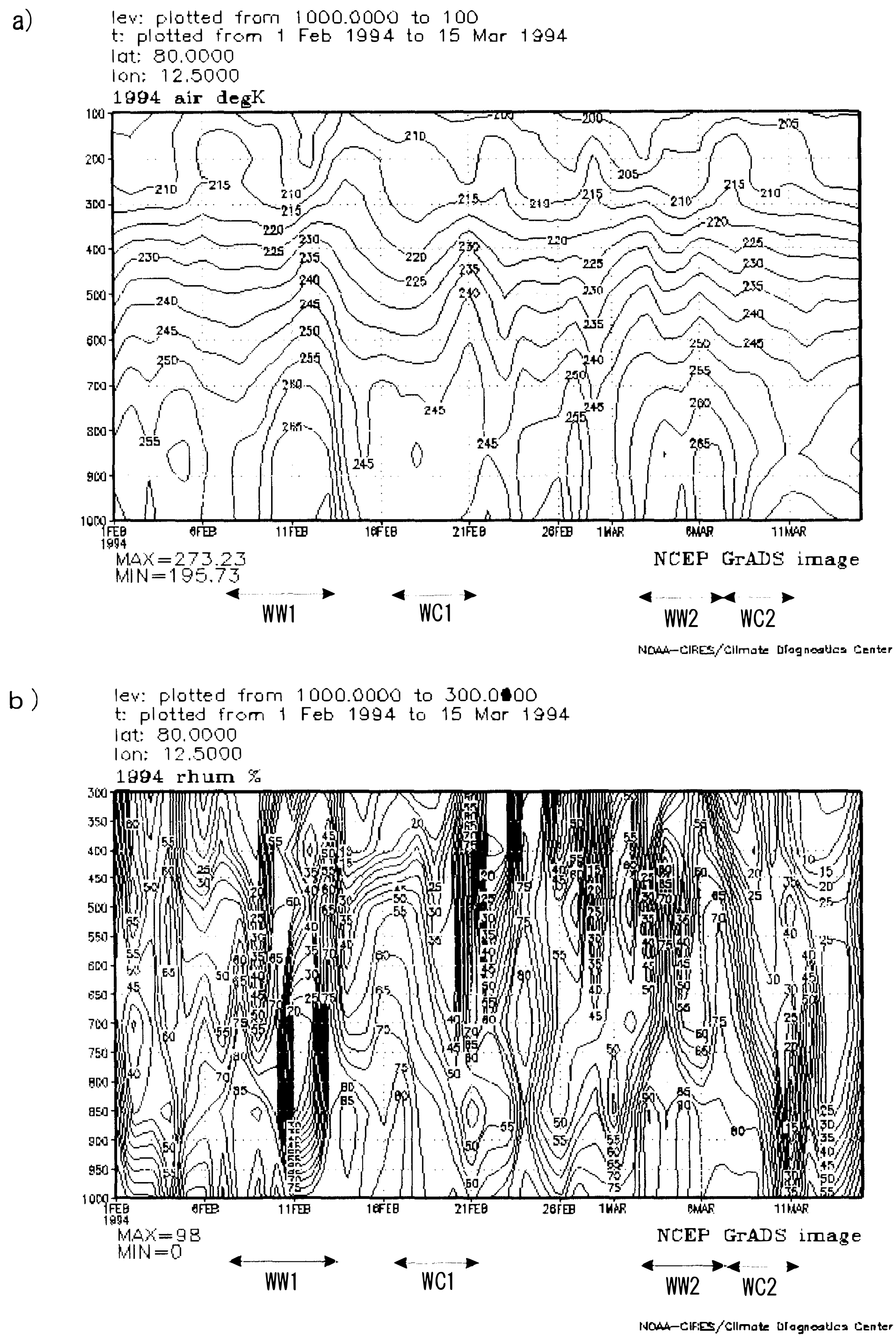

Fig. 6. Temperature and relative humidity at the grid point (80.0°N, 12.5°E) in winter 1994. a) Temperature. b) Relative humidity.

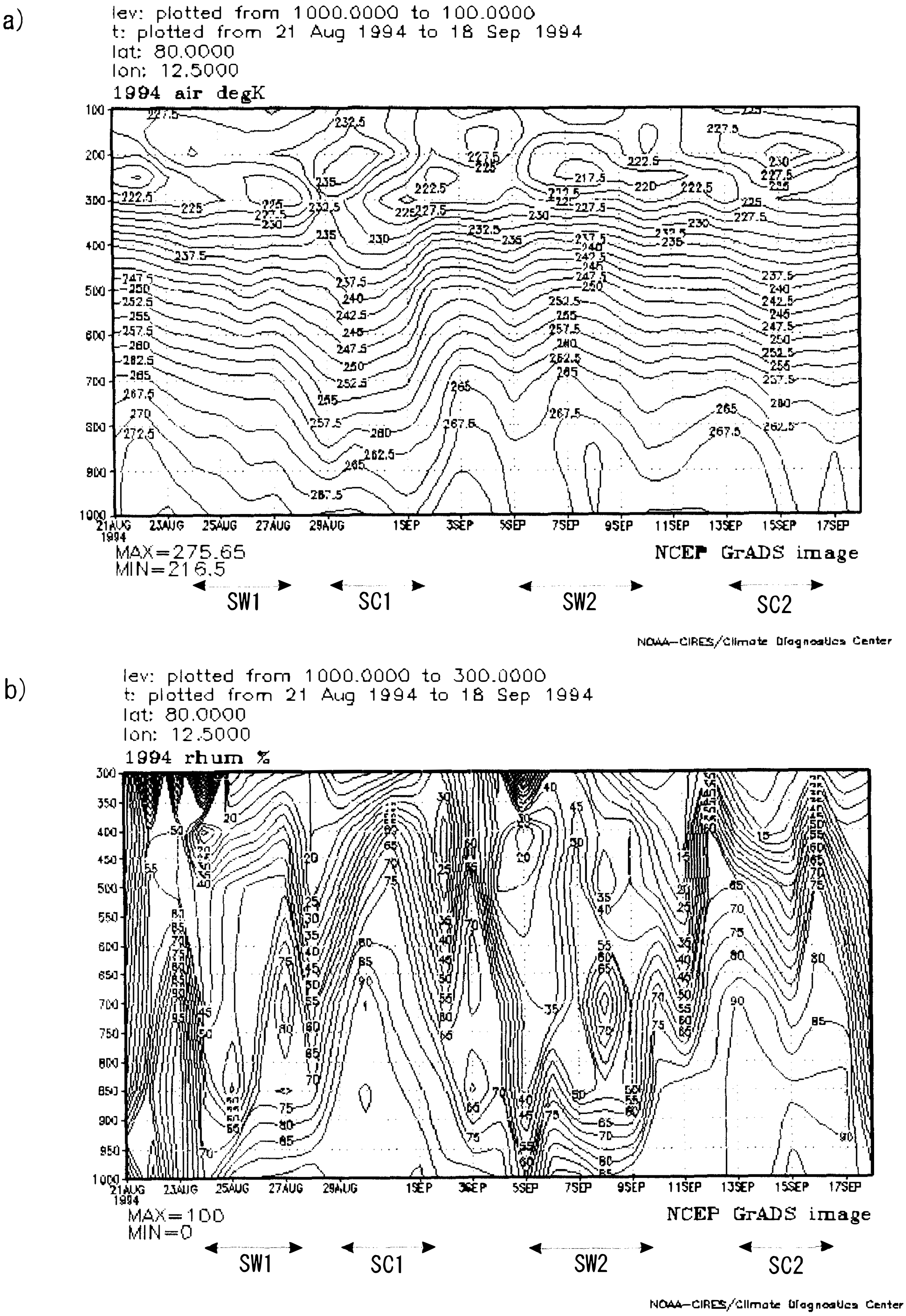

Fig. 7. Temperature and relative humidity of this grid point (80.0°N, 12.5°E) in summer 1994. a) Temperature. b) Relative humidity.

relative humidity at this grid point are shown in Fig. 6 (winter) and Fig. 7 (summer).

First, the relation between $\delta^{18}O$ and aerology data was examined in winter. Then, the following were divided: the period which the value of $\delta^{18}O$ was higher than the mean value (−12.9‰), and the period which it showed lower than that, as it was shown in Table 3. The periods in which two values were higher than the mean value continuously for several days were February 8–13 and March 2–7, and are called WW1 and WW2. The periods in which the value was lower than the mean value for several days were February 16–22 (except February 17; −11.7‰) and March 8–11, and are called WC1 and WC2.

In periods WW1 and WW2, there became the air masses which temperature were over 260 K from the surface to about 700 hPa (about 3000 m a.s.l.). In these two periods, there were the air masses which were the warmest throughout the observation period in February and March. Simultaneously, the values of the relative humidity of these warm air masses were over the 80%. These air masses apparently brought precipitation, because other wet air masses did not exist in WW1 and WW2.

In the meantime, temperatures of air masses between ground level and about 700 hPa were below 250 K in periods WC1 and WC2. The relative humidity in these periods were low value (below 70%), except for the layer over 850 hPa (below about 1400 m a.s.l.) became over 80% from February 16–17 and the layer between 700 hPa and 400 hPa (about 6500 m a.s.l.) became over 70% on February 21. Though in periods WC1 and WC2, the humidity was not very high, air masses in which temperature was almost constant existed between surface and a level of 700 hPa. It is likely that these air masses were inversion layer, and an ascending current was generated. In periods WC1 and WC2, the water vapor in the air was lifted from ground level into the sky by this ascending current, and precipitation occurred.

Properties of the air masses which brought about precipitation during WW1, WW2, WC1 and WC2 are collected in Table 3. From this table, the temperature of the air masses which brought about precipitation of WW1 and WW2 was higher at least 10 K than that of WC1 and WC2. The vertical extent of the moist air was also different between WW1, 2 and WC1, 2. In addition, amount of precipitation for WW1 and WW2 was over twice larger than that for WC1 and WC2. These results were suggested that fluctuation of $\delta^{18}O$ in the precipitation concerned with the characteristic of air masses which seemed to be supply source of the precipitation.

Then, the relation between $\delta^{18}O$ and aerology data was investigated in summer. The following were respectively extracted: Periods in which the $\delta^{18}O$ was higher than the mean value in summer (−12.3‰) for several days, are called SW1 (August 24–27) and SW2 (September 5–9), and periods in which low temperature persisted (SC1: August 29–September 1, SC2: September 13–16; *cf.* Table 3).

SW1 was one of the periods in which surface air temperature was highest during the observation period. In this period, the layer over 900 hPa (below 1000 m a.s.l.) was the highest for relative humidity (over 80%; Fig. 7b). The temperature between surface and a level of 900 hPa was over 267.5 K. In SW2, a layer which was similarly humid for a layer over 900 hPa existed. The temperature of this layer was over 265 K. Therefore, in SW1 and SW2 with high value of $\delta^{18}O$, the condensation of the water vapor which precipitated seemed to occur at low altitude and at high temperature.

In SC1 and SC2 where $\delta^{18}O$ was low, the layers between 600 hPa level (about 4000

Table 3. Properties of air masses which brought about precipitation and $\delta^{18}O$ during each period.

1993/94 Winter	WW1	WW2	WC1	WC2
Air masses				
Period	Feb. 8-13	Mar. 2-7	Feb. 16-22	Mar. 8-11
Vertical extent (hPa)	S*-700	S*-700	S*-850	S*-850
Temperature (K)	260-270	260-265	245-250	245-250
Precipitation (mm/half day)	0-34.9	0.1-8.1	0-3.2	0-0.8
$\delta^{18}O$				
Average (‰)	−9.8	−11.1	−16.6	−15.5
Maximum (‰)	−7.8	−7.3	−11.7	−13.7
Minimum (‰)	−12.8	−12.7	−20.4	−17.5

1994 Summer	SW1	SW2	SC1	SC2
Air masses				
Period	Aug. 24-27	Sep. 5-9	Aug. 29-Sep. 1	Sep. 13-16
Vertical extent (hPa)	S*-900	S*-900	S*-600	S*-600
Temperature (K)	267.5-275	265-270	245-270	250-267.5
Precipitation (mm/half day)	0.1-2.5	0-1.4	0.6-4.2	0.3-4.4
$\delta^{18}O$				
Average (‰)	−7.0	−6.8	−21.0	−17.8
Maximum (‰)	−5.8	−6.4	−18.5	−16.3
Minimum (‰)	−8.4	−7.1	−23.6	−19.1

*surface

m a.s.l.) and surface were high relative humidity, over 80%. Water vapor seemed to be condensing in these layer, since there was no other moist air. The temperature at the 600 hPa level was about 250 K in both periods.

The results in four summer periods are shown in Table 3. In SW1 and SW2, $\delta^{18}O$ clearly surpassed the mean value during the observation period, and the temperatures of the air masses which supplied the precipitation in these periods were over 265 K. $\delta^{18}O$ of SC1 and SC2 fell below the mean value, and the lowest temperature of the air masses which supplied the precipitation in these periods were below 250 K. The difference of the lowest temperatures between SW1, 2 and SC1, 2 was at least 15 K. Therefore, temperature of the air masses which supplied the precipitation seemed to be related to the change of the value of $\delta^{18}O$ in summer. Not only temperature but also amount of precipitation and extent of moist air were different between SW1, 2 and SC1, 2. Maximum Precipitations for half day in SW1 and SW2 were as almost half as that in SC1 and SC2. Moist airs in SW1 and SW2 existed only near surface, and in SC1 and SC2, spread from surface to 600 hPa level. It was found from these results that the characteristic of air masses brought precipitation influenced fluctuation of $\delta^{18}O$ in summer, same as winter.

4.3. Characteristic of variation of $\delta^{18}O$ in precipitations

Relationship between surface air temperatures and $\delta^{18}O$ in precipitations at Ny-Ålesund are appeared in Fig. 8. There is a week correlation between both in winter (R^2= 0.36; R^2 : Decision coefficient), and little in summer (R^2=0.22) as was suggested above.

In Table 3, we arranged the properties of air masses which brought about precipitation during 8 periods in both winter and summer. In winter, the temperature of the humid air

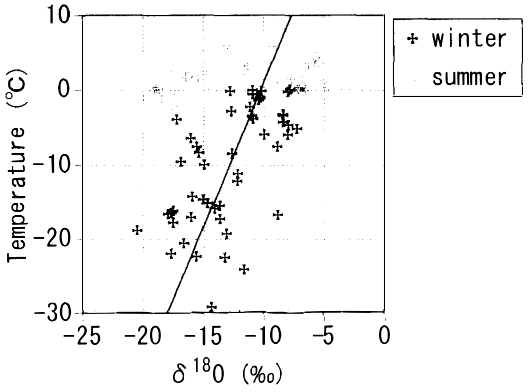

Fig. 8. The relationships between $\delta^{18}O$ and surface air temperature at Ny-Ålesund in winter and summer. The relationship in winter approximates the equation $T = 1.08\ \delta^{18}O + 3.63$ ($R^2 = 0.36$), and that summer the equation $T = 1.37\ \delta^{18}O - 3.52$ ($R^2 = 0.22$), T: surface air temperature, R^2: decision coefficient.

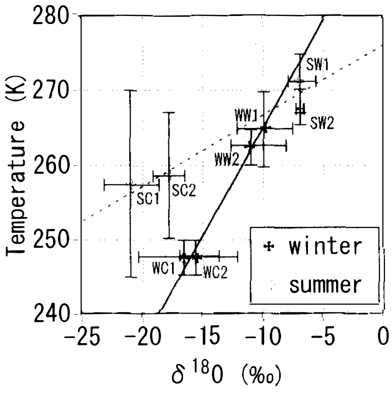

Fig. 9. The relationships between $\delta^{18}O$ and temperature of humid air masses which precipitations occurred in winter and summer. Vertical and horizontal bar means fluctuation range of temperature and $\delta^{18}O$ respectively. The relationship in winter approximates the equation $T = 2.82\ \delta^{18}O + 2.93 \times 10^2$ ($R^2 = 0.98$), and that summer the equation $T = 0.88\ \delta^{18}O + 2.75 \times 10^2$ ($R^2 = 0.93$), T: temperature of humid air mass, R^2: decision coefficient.

masses fluctuated between 245 K and 270 K. It fluctuated between 245 K and 275 K in summer. Though the maximum temperature was different by 5 K in winter and summer, the minima were equal. In Fig. 9, the relation between the temperature of the moist air masses and $\delta^{18}O$ in precipitations are shown, same as Fig. 8. Decision coefficient between both was 0.98 in winter and 0.93 in summer. Better correlations occurred between both in winter and summer. However there were the different equations which found the relationships between the temperature of the moist air masses and $\delta^{18}O$ in precipitations in winter and summer, as shown in Fig. 9, which possibly depended upon the atmospheric conditions in winter and summer. It means that $\delta^{18}O$ in precipitations at Ny-Ålesund in each season are approximated by these equations independently. From what has been discussed above, we can conclude that $\delta^{18}O$ in precipitations at Ny-Ålesund were decided by the temperature of the air mass in which the precipitation occurred.

A relation between $\delta^{18}O$ mean value of all precipitations during our observation and surface mean temperature at Ny-Ålesund agreed on proportional relationship of the other Arctic, as we mentioned in Section 4.1. It means that $\delta^{18}O$ during this observation was almost usual at Ny-Ålesund.

It has been reported that the seasonal variation of $\delta^{18}O$ was not recognizable in analysis of several ice cores from Svalbard, and the reason has been regarded that snow melted after it accumulated on the glacier in previous studies (Goto-Azuma *et al.*, 1995; Punning *et al.*, 1986. Besides, we could indicate other reason why the seasonal variation of $\delta^{18}O$ in precipitations did not recognize distinctly from this study.

5. Conclusions

Precipitation samples were collected at Ny-Ålesund in northwestern Svalbard in order to clarify the reason why the seasonal variation of $\delta^{18}O$ in ice cores from Svalbard was not so clear. We examined the relationship between $\delta^{18}O$ in precipitation and surface air temperature in winter and summer. Less correlation between $\delta^{18}O$ mean value in all precipitations during this observation and surface mean temperature was observed, not as it has previously been reported that there is generally a correlation between both in the polar regions. Therefore, it was indicated that present $\delta^{18}O$ observed value did not have the large difference from normal value. However, the extent of variation of $\delta^{18}O$ in winter and summer was similar in spite of a clearly difference of surface air temperature. It was concluded that in Ny-Ålesund, the fluctuation of $\delta^{18}O$ can not be explained only by the change of surface air temperature.

Then we examined the relationship between $\delta^{18}O$ change and aerology data. There was a correlation between temperature of the air masses which brought about the precipitation at Ny-Ålesund and $\delta^{18}O$ in both winter and summer. The relationships between both in winter and summer were found by the different equations, and $\delta^{18}O$ in each season are approximated by these equations independently.

In addition, we considered that it became a one of reason why the seasonal variation of $\delta^{18}O$ in analysis of several ice cores from Svalbard disappeared partly.

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