Isotopes in Precipitation in Northwestern North America

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ABSTRACT

Regional analyses of oxygen and deuterium isotope abundances in precipitation from selected stations of the International Atomic Energy Agency-World Meteorological Organization global netwok are presented as background for a more detailed study of the origin and history of atmospheric water in western Canada east of the Rocky Mountains. Departures from Dansgaard's regression between oxygen-18 concentrations and surface temperatures are attributed mainly to differences in intial water vapor isotope concentrations at inland stations in western Canada. Values of δO^{18} in rain at Edmonton are best correlated with 800 mb temperatures. However, snow data showed little variation in correlation with height up to 800 mb, and larger unexplained variance than rain, despite the fact that evaporation and isotope exchange effects are small for snow. Using simultaneous upper air temperature and wind data the δO^{18} variations in snow are attributed both to large condensation temperature variations that can occur in winter storms, and to the condensation of water vapor of different origin and history. Isotope concentrations characteristic of precipitation derived from re-evaporated water are associated with strong westerly flow below cloud level. On the other hand isotope concentrations characteristic of coastal station precipitation are associated with easterly low-level flow or light winds. The magnitude of the differences in concentrations in the two situations is so large that isotope measurements should be useful in the study of the structure of such storms. Previous studies have attributed the small slopes of δO^{18}-δD regression lines at Fort Smith, Bethel, and Whitehorse to the effects of rapid evaporation of precipitation below cloud base. An alternative approach using a fixed slope of 8 and varying intercepts suggests that the observed isotope concentrations can be accounted for by variations from near-equilibrium evaporation of North Pacific Ocean water in summer to rapid evaporation of the ocean water in winter. This explanation appears to be more consistent with the climate characteristics of these stations.

1. Introduction

Following the pioneering studies of Dansgaard (1953), Epstein and Mayeda (1953), Friedman (1953), and others, scientists in several branches of geophysics have sought improved understanding of fractionation processes for the most abundant isotopes of water H_2O^{16}, HDO^{16}, and H_2O^{18}. Fractionation of these isotopes in natural waters occurs mainly during condensation and evaporation processes and results in isotope mass-ratio changes that often exceed measurement errors by one to two orders of magnitude. Despite the appeal of stable isotope analysis in studies of the movement and transformation of water in the atmosphere and on earth, widespread use of the technique has not been realized in meteorology. It is true that interpretation of large-scale features of the hydrological cycle by this method alone is hampered by the complexity of fractionation processes and by the lack of adequate simultaneous measurements of temperature, humidity, and air trajectories. Nevertheless, when used together with other sources of information on sources and sinks of water and on fluxes of water vapor, it is believed that isotope analysis can be a valuable tool in diagnostic studies of atmospheric water.

The present study was motivated by interest in both the immediate and ultimate origin of water vapor and precipitation in the western prairie region of Canada. According to Rasmussen (1967) the water balance for this area is determined essentially by water vapor fluxes across the southern and western boundaries. Because of the mountain barrier to the west these fluxes exhibit a rather pronounced maximum above 850 mb. Mean vertical profiles of water vapor mixing ratios from radiosonde stations at Edmonton, Alberta, and Fort Smith, Northwest Territories, exhibit maxima between 800 mb and 850 mb in winter and suggest that local precipitation is derived principally from vapor imported from the North Pacific. However, from April to October inclusive, maximum mean mixing ratios are found at the surface with no evidence of

secondary maxima aloft. These profiles suggest that either local evaporation and transpiration or low-level advection from north or east are the immediate sources of vapor for precipitation. The orographic contribution associated with northeasterly upslope winds has been estimated by Reinelt (1970) to be not less than 13% of the total precipitation for the province of Alberta.

Improved knowledge of the sources of water vapor for precipitation can have immediate practical application. For example, if it is known that local evapotranspiration is the primary source in warm-season months, then one may expect the existence of a period of several weeks in spring, prior to the onset of significant transpiration, when local or distant melting snow is the only important source. This imposes an upper limit on the absolute humidity of the air in dry spells and provides a basis for better understanding of forest fire weather prediction problems at this time of year. Stable isotope concentrations in water vapor and precipitation represent an independent source of information on the origin and history of atmospheric water and for this reason a special sampling and analysis program was undertaken at the University of Alberta.

Comprehensive discussions of the physical processes that influence stable isotope ratios in precipitation have been given by Dansgaard (1964) and Craig and Gordon (1965). In addition, these authors have provided global analyses that serve as essential reference points for more local or regional water balance studies such as those of Dincer (1968), Gat and Carmi (1970), and Gat and Dansgaard (1972). First estimates of annual mean values and ranges of oxygen and deuterium isotope concentrations in precipitation for worldwide sampling stations included in the International Atomic Energy Agency (IAEA)—World Meteorological Organization (WMO) survey (IAEA 1969, 1970, 1971) were published by Dansgaard (1964). Annual mean values were mapped for the Northern Hemisphere by Dansgaard and Tauber (1969). Several basic results have emerged from these and many other studies. Following Dansgaard (1964) and Dincer (1968) some of these results are:

1) The isotopic composition of ocean water is nearly constant with concentrations of 997, 680:2000:320 ppm for H_2O^{16}, H_2O^{18}, and HDO^{16}, respectively.
2) Because of differences in transport rates to and from liquid water surfaces, the isotopic species are fractionated during evaporation and condensation processes.
3) Under equilibrium conditions the fractionation factors for evaporation are given by the ratios of the vapor pressure of the light isotope to the vapor pressure of the heavy isotope. These ratios are temperature dependent.
4) Precipitation becomes isotopically lighter with increasing altitude and latitude and from warm to cold season. This variation is quite well correlated with air temperature.
5) The concentrations of H_2O^{18} aud HDO^{16} in water are linearly related to each other within the usual range of atmospheric temperature. However, the constant of proportionality varies depending on atmospheric humidity, kinetic processes during evaporation, and isotopic exchange processes.

2. Experimental program

A water sampling and oxygen isotope analysis program was initiated by the Institute of Earth and Planetary Physics at the University of Alberta in 1970. Several hundred samples of rain, snow, water vapor, and lakewater were collected over a three-year period. On several occasions sequential precipitation samples were obtained at one or more points within individual rain or snow storms. Samples (40 ml) were collected in open pans, transferred to glass bottles, and sealed during storms. Laboratory analysis procedures were similar to those described by Epstein and Mayeda (1953). A modified adjustable gas inlet system as described by Coleman and Gray (1972) was used. Reproducibility was better than 0.3‰. Water vapor samples were collected by drawing air slowly through a coil embedded in a slushed CO_2-ethanol mixture. Laboratory tests confirmed that nearly complete recovery of vapor was accomplished by this method. Reproducibility for water vapor samples was found to be ±0.6‰. Unfortunately, no facilities were available for analysis of deuterium content.

3. Analysis of monthly mean precipitation samples

Three to six years of data on oxygen isotope and deuterium contents in monthly precipitation samples are now available for most stations in the IAEA-WMO network (IAEA 1969, 1970, 1971). Estimates of annual mean values and variability were computed for stations in western Canada and for neighboring stations in the United States (Fig. 1). Missing data, including months without precipitation, occurred infrequently. Such gaps were filled by mean values computed from all available data for the month in question. All isotope data are represented as per mil (‰) deviations from an arbitrary standard called SMOW (Standard Mean Ocean Water) as defined by Craig (1961a). The notations δO^{18} and δD are defined by

$$\delta = \frac{R - R_{\mathrm{SMOW}}}{R_{\mathrm{SMOW}}} \times 10^3‰,$$

where the mass ratio $R = O^{18}/O^{16}$ or D/H, respectively. The precision of measurements is ±2‰ for δD and ±0.2‰ for δO^{18}. The results are listed in Table 1.

It is of interest to note that year-to-year variations in δO^{18} and δD as measured by the standard deviations

FIG. 1. Locations of isotope sampling stations used in this study.

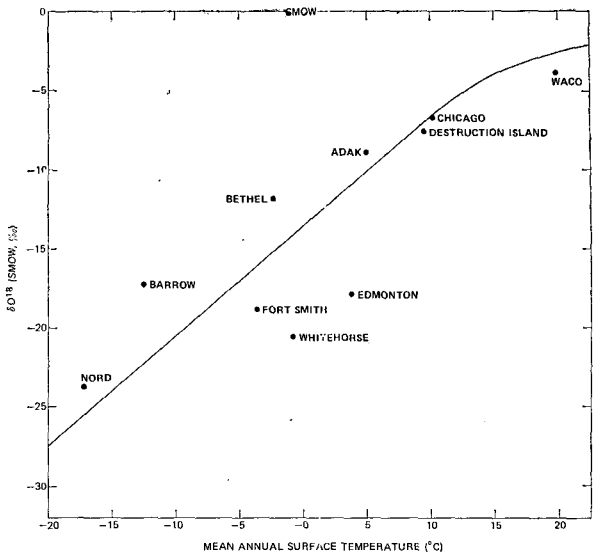

FIG. 2. Mean annual δO^{18} values in relation to mean annual surface temperature. Solid line from Dansgaard (1964).

in Table 1 often greatly exceeded the long-term precision limits. Furthermore, the departures from mean values in particular years seemed to be systematic. In 1964, for example, all stations except Barrow and Waco exhibited above average δO^{18} values with a mean departure of $+0.8‰$. In 1965 all stations except Whitehorse exhibited below-average departures with a mean departure of $-0.7‰$. It is unfortunate that δO^{18} and δD data collection for the IAEA-WMO network appears to have been terminated in 1966 or 1967 for all stations in western North America. Because of the dependence of δO^{18} and δO values on evaporation rates and, therefore, on sea-surface temperatures, studies of annual anomalies over periods of several years may be helpful in clarifying problems of weather pattern teleconnections (Bjerknes, 1969).

The following linear regression equation between mean annual surface temperature (T,°C) and δO^{18} values was derived by Dansgaard (1964) using coastal and island stations in tropical and middle latitudes, and high altitude station data from Greenland and Antarctica.

$$\delta O^{18} = 0.69T + 13.6‰. \quad (1)$$

The excellence of the regression led Dansgaard to conclude that the effective condensation temperature of water vapor must vary in parallel with the surface temperature. Significant departures from the regression can be expected in the following circumstances:

1) If precipitation occurs in air that is significantly more stable than the nearly moist-adiabatic condition found for low-latitude maritime stations the corresponding surface temperature will be low and annual mean δO^{18} values will fall above the regression line (1) above.
2) The effects of increased stability can be reduced or even reversed as station altitude increases to and beyond the mean condensation level.
3) If the effective condensation level is higher and colder than normal, such as might be expected at stations with heavy summer convective precipitation and light winter precipitation, the corresponding surface temperature will be high and the mean annual δO^{18} value may fall below the regression line (1).
4) If local precipitation is derived from condensation of water vapor produced by re-evaporation of precipitation, i.e., vapor having a low initial δO^{18} not in equilibrium with SMOW, the mean annual δO^{18} value may fall below the regression line (1) by as much as 10‰ in the case of complete re-evaporation of precipitated water.

TABLE 1. Precipitation-weighted annual mean δO^{18} and δD values for selected stations.

Station	No. of yrs	Temp. (C)	Precip. (mm)	δO^{18} (‰)		δD (‰)	
Waco	4	19.9	630	−3.9	±1.3	−23.0	±6.0
Chicago	6	10.2	830	−6.6	±0.9	−44.4	±6.0
Destruction Island*	4	9.5	2020	−7.6	±0.8	−53.4	±8.3
Edmonton	4	3.8	440	−17.9	±0.9	−137.3	±5.5
Fort Smith	4	−3.7	350	−18.8	±0.4	−146.7	±2.2
Whitehorse	4	−0.9	270	−20.6	±0.8	−159.2	±3.7
Bethel	4	−2.4	360	−11.9	±1.3	−94.5	±7.5
Barrow	3	−12.5	130	−17.2	±0.6	−132.2	±5.4
Adak Island	4	5.0	1460	−8.8	±1.0	−63.8	±6.2
Nord	5	−17.3	160	−23.7	±1.1	−178.1	±8.2

* Temperature and precipitation data from Tatoosh Island.

The mean annual δO^{18} values in Table 1 are compared with Dansgaard's regression line in Fig. 2. The

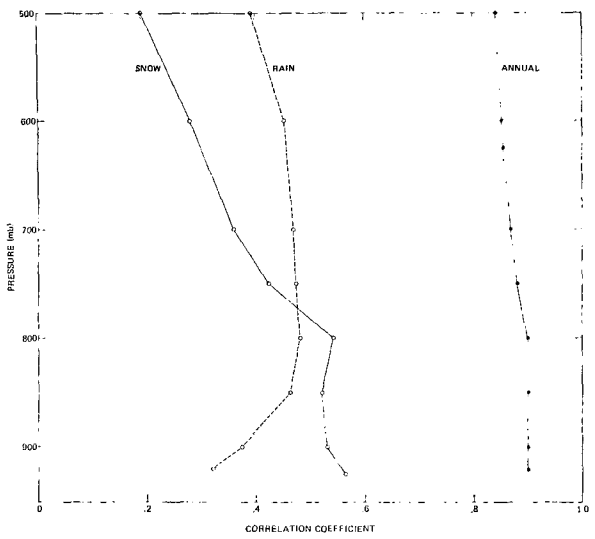

FIG. 3. Correlation coefficients between δO^{18} values in precipitation and air temperature at Edmonton 1970–72.

effects of stability are shown by Barrow, Bethel, and Adak Island, all high latitude stations near sea level. On the other hand the effects of stability at interior continental stations such as Edmonton, Fort Smith, and Whitehorse were more than compensated for by the effects of altitude and re-evaporation of precipitation. The effects of changes in condensation level on mean annual δO^{18} values appear to be less important than other factors as shown by the similarity of Edmonton, with a pronounced summer maximum of convective precipitation, and Whitehorse, with a winter maximum of precipitation.

The effects of stability and changes in effective condensation level are more clearly illustrated by means of upper air temperature data. Precipitation samples collected at Edmonton in 1970–72 were correlated with simultaneous upper air temperatures interpolated in time from radiosonde data at 12-hour intervals at Stony Plain, about 50 km west of the city. Height variations in correlation coefficients in rain (April to September) and in snow (October to March) are shown in Fig. 3. The maximum correlation in rain occurred at 800 mb. Surprisingly the correlation between δO^{18} and temperature in snow was nearly constant from the surface to 800 mb. Mean vertical temperature profiles for all rain and snow samples are shown in Fig. 4. Because 70% of the total precipitation at Edmonton is rain it is evident from Fig. 4 that stability and related altitude effects on surface temperature—δO^{18} regression are not very pronounced at that location.

TABLE 2. Temperature $-\delta O^{18}$ regression statistics for Edmonton.

Season	Number of samples	Total variance $(‰)^2$	800-mb temperature regression: Correlation coefficient	800-mb temperature regression: Residual variance $(‰)^2$
Summer (rain)	133	9.8	0.48	7.6
Winter (snow)	72	14.0	0.54	9.8
Annual	205	53.7	0.90	10.2

Temperature—δO^{18} regression statistics for Edmonton are summarized in Table 2. It is clear from Table 2 that a large fraction of the total variance of δO^{18} readings in precipitation is not accounted for by a simple correlation with air temperature. It is known that fractionation due to evaporation of precipitation as it falls and vapor-liquid exchange processes contribute to some of this variability. However, it is surprising to find that both the total variance and the residual variance in δO^{18} are larger for snow than for rain in view of the fact that evaporation and exchange processes are negligible for the former. An example of the variability in δO^{18} in snow is shown in Fig. 5. The tendency for δO^{18} to rise with time was fairly typical but examples of the reverse trend were also found. The large varia-

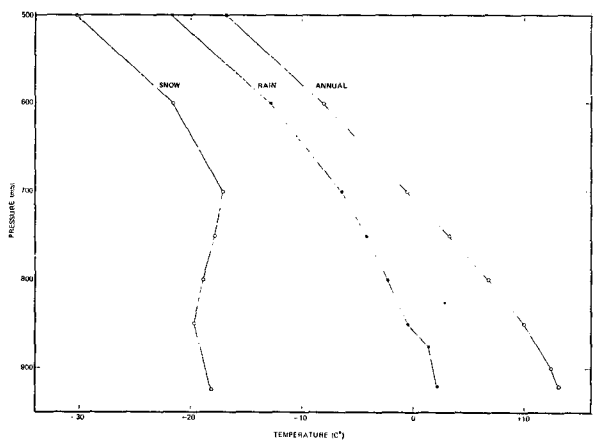

FIG. 4. Mean vertical profiles of air temperature in precipitation for sampling times used in Fig. 3.

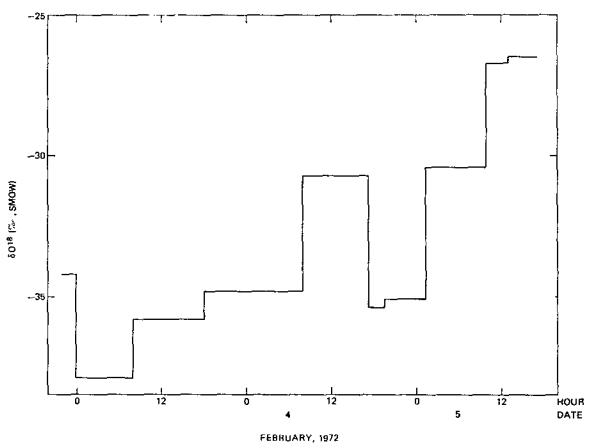

FIG. 5. Variations of δO^{18} with time in snow, 3–5 February 1972, at Edmonton.

TABLE 3. Linear regression statistics for δD-δO^{18} data for selected stations in western North America.

Station	Slope			Annual values		
	Summer	Winter	Annual	Intercept	Standard error	Correlation coefficient
Chicago	5.7	7.1	6.9	+0.9	6.1	0.98
Destruction Is.	7.6	7.5	7.2	−0.2	4.1	0.97
Edmonton	7.8	7.8	7.7	+0.4	6.2	0.99
Fort Smith	5.9	7.0	6.8	−20.9	7.3	0.98
Whitehorse	6.0	5.7	6.1	−34.2	8.9	0.95
Bethel	5.5	6.4	5.9	−25.2	5.9	0.97
Barrow	7.6	7.9	7.4	−2.9	7.2	0.98
Adak Island	6.3	7.2	6.8	−3.2	5.7	0.92

bility of isotope concentrations in snow is examined in more detail in Section 5.

4. Deuterium-oxygen 18 relationships

On the basis of 400 samples of water from streams, lakes, and precipitation, Craig (1961b) derived the following linear relation between δD and δO^{18}:

$$\delta D = 8\delta O^{18} + 10. \quad (2)$$

According to Craig the slope and intercept in (2) can be accounted for by kinetic evaporation of sea water ($\delta D/\delta O^{18} \simeq 5$) followed by condensation and precipitation in poleward moving air ($\delta D/\delta O^{18} = 8$). Regression lines for all stations except Nord listed in Table 1 were found to be

$$\delta O^{18} < -3: \quad \delta D = 8.0\delta O^{18} + 5.5 \pm 6.7 \quad (r = 0.99) \quad (3)$$

$$\delta O^{18} > -3: \quad \delta D = 5.4\delta O^{18} - 0.9 \pm 8.0 \quad (r = 0.62). \quad (4)$$

These results are in close agreement with Craig's findings except for the intercept in (3). Equation (3) was derived from monthly means for all available years for all stations (106 cases). Equation (4) was derived from individual monthly values (34 cases) mostly from Waco and Chicago.

Dansgaard (1964) using least-squares regression analysis on monthly values from individual stations found slopes that varied from 3.5 to 8.0 with intercepts of −30 to +30. These calculations were repeated for almost all stations in Table 1 using the larger sample of data now available. The results (Table 3) were in close agreement with Dansgaard's earlier findings. However, this approach to the analysis and interpretation of δD–δO^{18} data leads to some apparent incon-

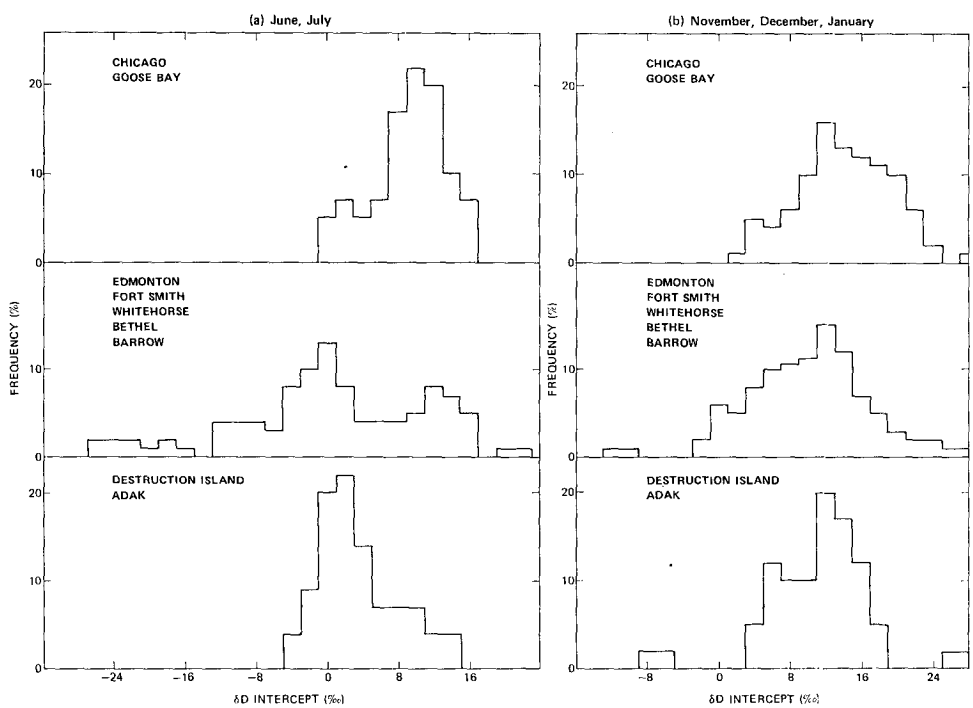

FIG. 6. Frequency distributions of intercept x in $\delta D = 8\delta O^{18} + x$ for selected groups of stations in summer and winter.

sistencies. For example, according to Dansgaard (1964), Fort Smith, Northwest Territories, with a warm season slope of 5.9 (Table 3) experiences rapid evaporation of precipitation while Edmonton with a slope near 8 does not. This is at variance with climatic data, which show Edmonton to be both warmer and drier than Fort Smith. Plots of monthly ratios on a $\delta D - \delta O^{18}$ diagram show that most high latitude stations exhibit a seasonal cycle from values near the line $\delta D = 8\delta O^{18} + 10$ in winter to values near $\delta D = 8\delta O^{18}$ in summer. Therefore, an alternative approach to the problem is to assume a constant slope of 8 for middle and high latitude stations and to compute variations in intercept alone. The results provide an alternative interpretation of $\delta D - \delta O^{18}$ data which is more consistent with the climatic characteristics of these stations and which accounts for the small apparent slopes at Fort Smith, Bethel, and Whitehorse in Table 3.

Figure 6 shows frequency distributions of δD intercepts x in the equation $\delta D = 8\delta O^{18} + x$ for different months and stations. The frequencies were smoothed by choosing overlapping intercept classes of size 6‰. The top and bottom station groups in Fig. 6 represent a western Atlantic and Gulf of Mexico source region (Chicago, Goose Bay) and a northeast Pacific source region (Destruction Island, Adak Island). Values of $\delta O^{18} > -3.0$ were not used for reasons described earlier. Chicago and Goose Bay were found to be very similar in all months for which $\delta O^{18} < -3$. The most frequent intercept at both stations was +10 to +12‰. Adak Island and Destruction Island on the other hand oscillated between a modal intercept of 0 to +2 in June and July to +12 in November, December, and January. Stations in western Canada and Alaska (grouped because of relatively small sample sizes) fell between these limits with a bimodal distribution in summer (modes of 0 and +12) and a broad distribution with maximum frequency at +12 in winter. As an alternative to Dansgaard's explanation of the small $\delta D/\delta O^{18}$ slope at Fort Smith, Bethel, and Whitehorse, it is suggested that much, though by no means all, of the summer precipitation at these stations results from near-equilibrium evaporation of North Pacific sea water followed approximately by Rayleigh condensation ($\delta D = 8\delta O^{18}$). On the other hand, winter precipitation from the North Pacific, and precipitation in all seasons from the Western Atlantic, corresponds more closely to rapid evaporation of sea water followed by condensation ($\delta D = 8\delta O^{18} + 12$). It must be noted that this explanation does not account for the scatter of negative intercepts at western Canadian and Alaskan stations. Although an intercept of +12 was most common at Chicago and Goose Bay in winter, the distribution was rather broad and intercepts as high as +20 were not uncommon. Such a feature could be accounted for by large month-to-month variations in the amount of kinetic evaporation. Assuming independent measurement errors of ±0.2‰ in δO^{18} and ±2.0‰ in δD, the expected standard deviation of intercept due to measurement errors is only ±2.6‰. It is assumed in the above analysis that western Atlantic and North Pacific sea water do not deviate appreciably from SMOW and this is supported by data presented by Craig and Gordon (1965). Although the data seem to suggest a process of near-equilibrium evaporation of sea water in summer, it must be emphasized that this is a first approximation only and that true equilibrium evaporation may not exist in nature (Craig and Gordon, 1965).

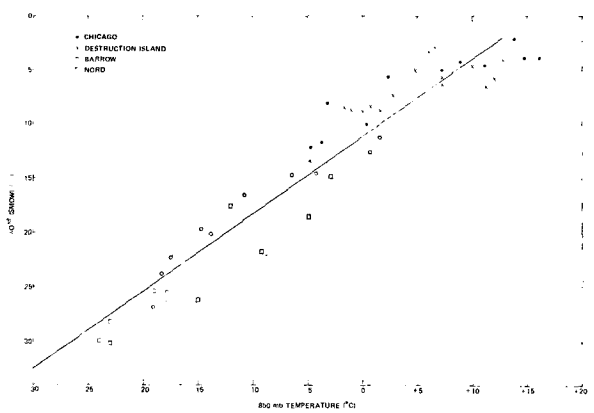

FIG. 7. Monthly mean δO^{18}-850 mb temperature scatter diagram and least-squares regression line.

5. Oxygen isotope variations in snow

In Section 3 it was shown that low-level stability changes and interrelated altitude effects on δO^{18}-annual mean temperature regressions were not very pronounced at Edmonton. Furthermore, a comparison of Edmonton data (summer precipitation maximum) with Whitehorse data (winter precipitation maximum) showed that changes in mean condensation levels associated with convection were much less important than other factors. In order to investigate δO^{18} changes in snow it is necessary to consider monthly or seasonal rather than annual values. A scatter diagram showing monthly mean δO^{18} readings in relation to simultaneous monthly mean 850 mb temperatures for coastal stations Destruction Island, Barrow, and Nord, and for an inland station at fairly low latitudes (Chicago) is given in Fig. 7. Each point represents a monthly mean for all available years in the IAEA-WMO survey. The regression line in Fig. 7 is

$$\delta O^{18} = 0.71T - 11.1 \pm 2.4‰. \quad (2)$$

where T is mean monthly 850 mb temperature (°C). The slope of (2) is not significantly different from that derived by Dansgaard (1964) and given by (1) for mean annual values. The scatter in Fig. 7 is large as would be expected because of factors listed previously. However, there seems to be no systematic departure from the population represented by the regression line.

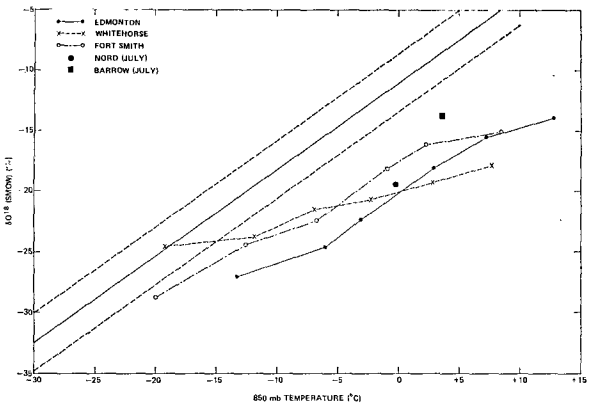

FIG. 8. Mean δO^{18} values in 5°C temperature ranges for Edmonton, Whitehorse, and Fort Smith with regression line from Fig. 7.

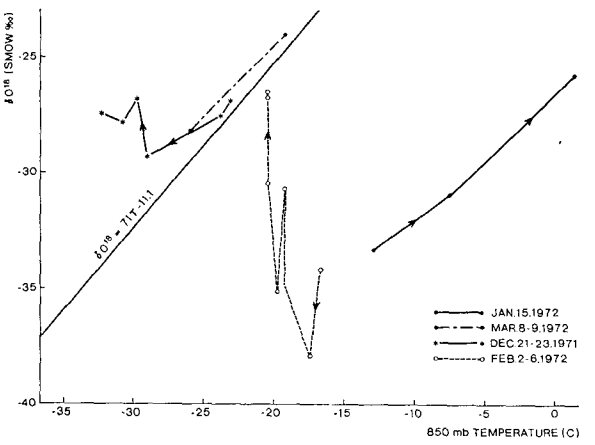

FIG. 10. Values of δO^{18} and 850 mb temperatures for individual snow samples in 4 storms at Edmonton.

The regression line (2) and broken lines representing deviations of one standard error unit are reproduced in Fig. 8 along with July means for Nord and Barrow (not included in Fig. 7), and mean δO^{18} values for 5°C temperature ranges at Edmonton, Whitehorse, and Fort Smith. Evidently warm season precipitation at the latter stations and at Nord and Barrow in July had δO^{18} values about 10‰ below those expected at coastal stations even at high latitudes. Similar behavior was exhibited by short-period precipitation samples collected in Edmonton in 1970–72. Average δO^{18} values and their variability (± 1 standard deviation) in 10°C temperature intervals for these samples are shown in Fig. 9.

On the basis of the results shown in Figs. 8 and 9 it is argued that almost all warm season precipitation at Edmonton, Whitehorse, and Fort Smith, shielded by high mountain ranges to the west, originated from water vapor produced by re-evaporation of precipitation over land (including transpiration). At low temperatures, however, precipitation at these stations

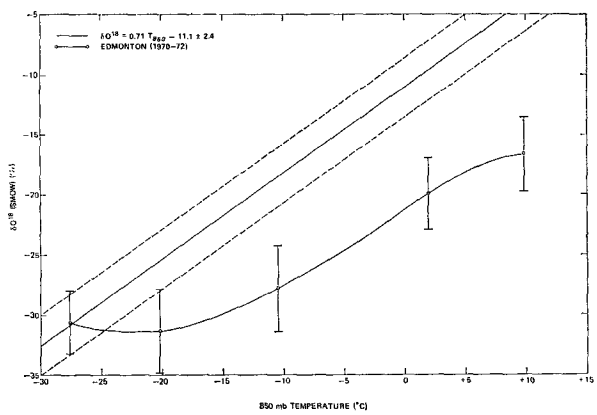

FIG. 9. Mean δO^{18} values in 10°C temperature ranges for short-period precipitation samples at Edmonton (1970–72). Vertical bars mark ± 1 standard deviation.

originated mainly from directly imported ocean vapor just as it does all months apparently at Chicago and Destruction Island.

According to Figs. 8 and 9 the regression lines representing these two water vapor sources come together at low temperatures. However, the δO^{18} variations in individual snowstorms are masked here by the statistical analysis. Figure 10, for example, shows δO^{18} variations in relation to 850 mb temperatures in four separate snowstorms at Edmonton and provides an explanation for much of the observed large variability of δO^{18} in snow. Evidently, some snowstorms are characterized by δO^{18}—850 mb temperatures typical of re-evaporated precipitation (15 January 1972), others by values typical of precipitation from ocean vapor (21–23 December 1971 and 8–9 March 1972), and still others involve transitions from one vapor source to the other (2–6 February 1972). In other words, the large δO^{18} variations in snow at Edmonton were caused by

a) large time variations in condensation temperature within single storms, and
b) participation of water vapor of different origin and history within the same storm.

Additional support for (b) was found in the observed wind field. Resultant winds for the surface—800 mb layer for the 12-hour period preceding sample collection time were evaluated for each day. In order to eliminate changes due to variations in condensation temperatures all δO^{18} values were expressed as deviations Δ from the regression line (2) reduced to a temperature of −20°C. Using (2),

$$\Delta = 0.71T_o - \delta_o O^{18} + 25.3, \qquad (3)$$

where T_o and $\delta_0 O^{18}$ are the observed 850 mb temperature and δO^{18} value for each sample. Calculated deviations, including standard deviations for days with multiple samples, are plotted at the origins of the resultant wind vectors in Fig. 11. It will be seen that

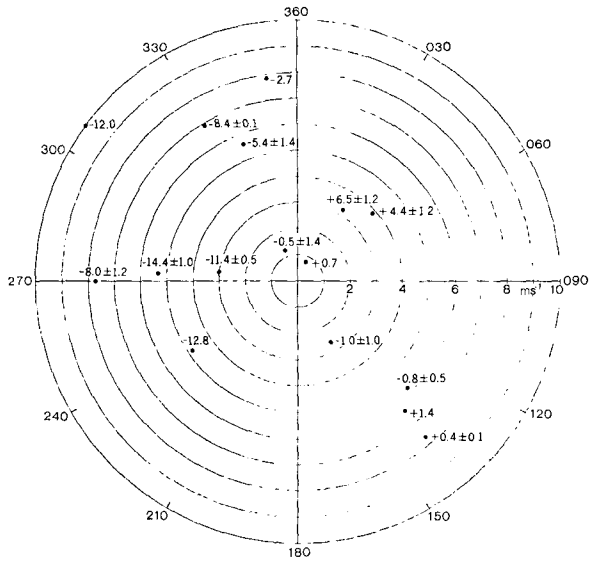

FIG. 11. Values of δO^{18} (per mil) in snow, expressed as deviations from the regression line $\delta O^{18}=0.71T-11.1$ reduced to a common 850 mb temperature of -20°C and plotted at the origins of 12-hour resultant winds for the surface-800 mb layer at Edmonton.

large negative deviations, i.e., δO^{18} values near the warm season regression line for Edmonton, were associated with westerly winds, while small positive or negative deviations, i.e., δO^{18} values near the coastal station regression line were associated with easterly winds or very light winds. The implication of Figs. 10 and 11 appears to be that in Alberta in winter, when surface evaporation and transpiration are negligible over a large surrounding area, δO^{18} values is snow are sensitive indicators of whether or not water vapor has crossed the Rocky Mountain barrier after experiencing at least one evaporation-precipitation-re-evaporation cycle before participating in the local storm.

6. Oxygen isotope variations at high temperatures

The tendency for δO^{18} values to approach a constant value at high temperatures as shown by Dansgaard (1964), by the monthly mean values for Chicago and Destruction Island (Fig. 3), and by the short-period samples from Edmonton (Fig. 4), is examined more closely in Fig. 12. Mean values of δO^{18} in 5°C temperature ranges clearly show this trend in both sets of observations. At the highest temperatures the frequency distribution of δO^{18} values became markedly skewed (Fig. 12, right side). In other words, as one approaches high surface or 850 mb temperatures the effective condensation temperature no longer varies in parallel with low-level temperatures. In Alberta, precipitation at these temperatures is largely convective in origin. One is forced to conclude that 850 mb temperatures measured at 12-hour intervals at the radiosonde station 50 km west of Edmonton are not well correlated with effective condensation temperatures in showers and thunderstorms over the city.

7. Summary

Data from the International Atomic Energy Agency —World Meteorological Organization global survey of stable isotopes in precipitation together with more recent oxygen isotope measurements in precipitation in central Alberta were used to derive estimates of annual means and variability of δO^{18} and δD for stations in western North America as background for more detailed studies of the origin and history of water vapor in the lee of the Canadian Rocky Mountains. Departures from Dansgaard's (1964) regression between δO^{18} and surface temperature were attributed to stability effects at Alaskan stations near sea level. However, differences in initial vapor contents of oxygen-18, such as those caused by re-evaporation of precipita-

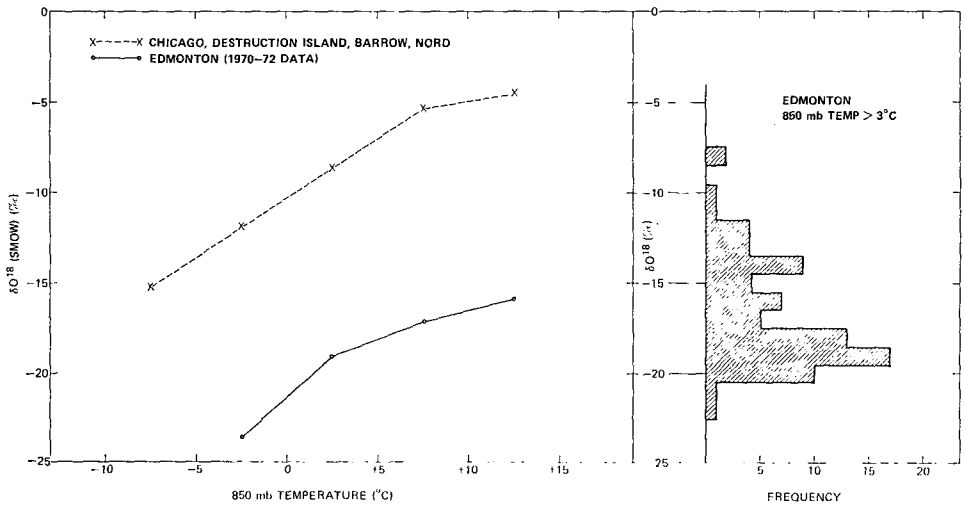

FIG. 12. Mean δO^{18} values in 5°C temperature ranges at high temperatures showing systematic departures from linear regression.

tion, appeared to be more important at inland stations in northwestern Canada. Values of δO^{18} in rain at Edmonton were best correlated with 800 mb temperatures. However, δO^{18}-air temperature correlations in snow were nearly constant from the surface to 800 mb. Large variations in δO^{18} (10 per mil or more) were found within and between individual snowstorms. These variations were attributed to large changes in effective condensation temperatures and to transitions from snow produced from water vapor having an upslope trajectory from the east to snow produced from water vapor that had crossed the Rocky Mountains to the west. The latter mechanism offers interesting possibilities for the use of oxygen isotopes as tracers in the study of snowstorm structure.

Regional deviations from Craig's (1961b) regression between oxygen and deuterium isotope ratios were studied using global survey data. Seasonal variations from quasi-equilibrium evaporation of North Pacific sea water in summer to rapid evaporation of sea water in winter was offered as an alternative to Dansgaard's (1964) explanation for $\delta D/\delta O^{18}$ ratios at Fort Smith, Bethel, and Whitehorse. Such an explanation appears to be more consistent with the climatic characteristics of these stations.

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