Geochemical Journal, Vol. 37, pp. 519 to 530, 2003

Seasonal variation in stable isotopic composition of alas lake water near Yakutsk, Eastern Siberia

KIMPEI ICHIYANAGI,[1]* ATSUKO SUGIMOTO,[2] ATUSI NUMAGUTI,[1,3] NAOYUKI KURITA,[1] YOSHIYUKI ISHII[4] and TETSUO OHATA[1,4]

[1]Frontier Observational Research System for Global Change, Syouwa-machi 3173-25, Kanazawa-ku, Yokohama, Kanagawa 236-0001, Japan
[2]Center for Ecological Research, Kyoto University, Kamitanakami Hirano-cho, Otsu 520-2113, Japan
[3]Graduate School of Environmental Earth Science, Hokkaido University, Kita 10 Nishi 5, Kita-ku, Sapporo 060-0810, Japan
[4]Institute for Low Temperature Research, Hokkaido University, Kita-19, Nishi-8, Kita-ku, Sapporo 060-0819, Japan

(*Received September 27, 2002; Accepted April 1, 2003*)

Detailed measurements of temporal variations in the stable isotopic composition of precipitation and lake water were conducted in the permafrost region near Yakutsk, eastern Siberia. The $\delta^{18}O$ ranged from approximately –30 to –5‰ in precipitation and from –25 to –5‰ in lake water were observed from April to August, 2000. Temporal changes in $\delta^{18}O$ of precipitation observed weekly at 12 sites all showed the same trend. The temporal variation of $\delta^{18}O$ in lakes classified into two groups: isotopically steady-state lakes with less than 5‰ variation, and non-steady-state lakes with variation exceeding 10‰. In non-steady-state lakes, the water originated from snowmelt, and the $\delta^{18}O$ of lake water gradually enriched as a result of evaporation during the summer. In steady-state lakes, the water originated predominantly from ^{18}O-enriched lake water that had evaporated in the previous summer. The temporal volumetric and isotopic variations in alas lakes are accurately depicted by an isotope mass-balance model using Rayleigh fractionation over daily time steps. The inflow of soil water (subsurface flow) was estimated to be constant (200 m^3/day) from May to August, based on the difference between observed and simulated lake volumes. Taking the isotopic mass-balance into consideration, the soil water in lower part of the active layer is inferred as a major component of inflow which has a $\delta^{18}O$ of about –23‰.

INTRODUCTION

Thermokarst landforms are abundant in the permafrost regions of eastern Siberia, because large seasonal changes in soil temperature promote thermal erosion (French, 1984). Alasses are thaw-depressions that range from a few hundred meters to about 10 km across, which occupy about 30% of the land-surface area on the right bank of the Lena River (Suzuki *et al.*, 2001). In areas with a high permafrost ice content, alasses commonly form thermokarst lakes, hereinafter it called “alas lakes”. Because stream networks are not present and runoff is minimal except for snowmelt, evaporation from lake surfaces is the dominant outflow in alas lakes (Ishii and Yabuki, 2001). In permafrost regions, alas lakes have an important role in the hydrological cycle.

There have been some studies on the water cycle of thermokarst lakes in North America (e.g., Gibson, 2002), but little research has been done in Siberia. Although climatic conditions in Siberia are more amenable to thermokarst lake development than those in North America (French,

**Corresponding author (e-mail: kimpei@jamstec.go.jp)*

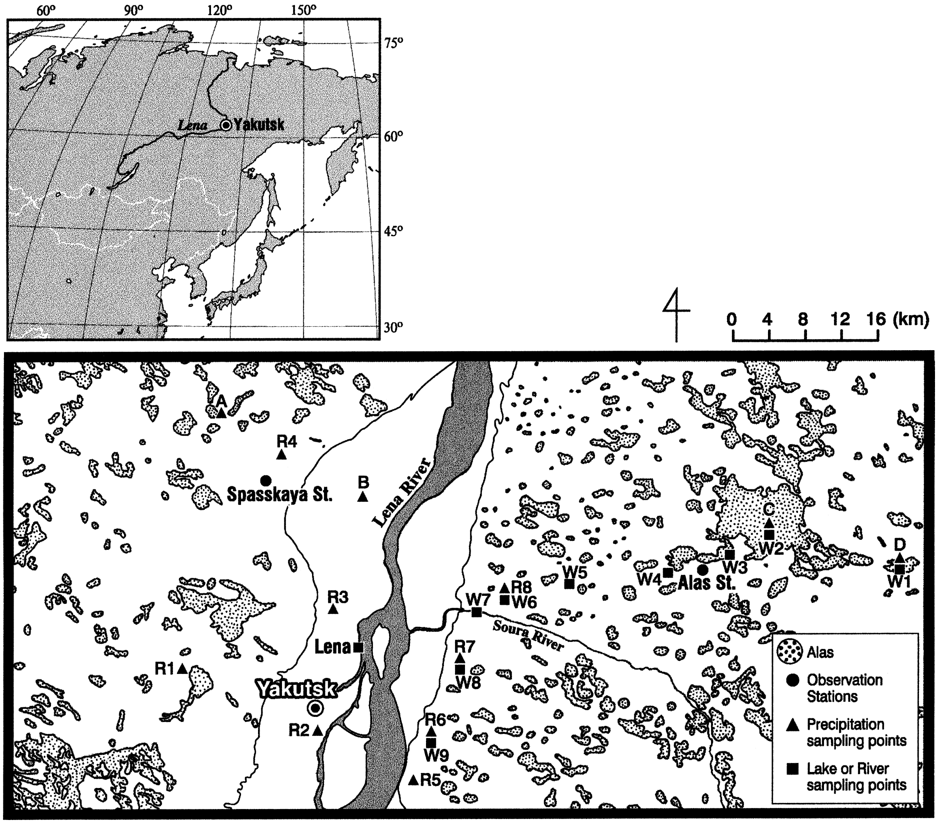

Fig. 1. Locations of precipitation and alas lake water sample sites.

1984), there is insufficient information on the hydrological cycle in Siberia's permafrost region. In recent studies, large-scale water vapor transport has been considered using stable isotopes in precipitation (e.g., Ichiyanagi *et al.*, 2002). Observed isotopic data is limited in high latitude, so it is important to validate the results of Atmospheric General Circulation Model (AGCM).

Soil water volumes were high in spring 2000 (Sugimoto *et al.*, 2003), and the volume of alas lakes and river runoff has been extremely high for the last 50 years (Desyatkin and Desyatkin, 2001). This study examines the seasonal variation of stable isotopic compositions in precipitation and surface water, and interprets of the origin of alas lake water. The volumetric and isotopic temporal variation of alas lake is estimated using an isotopic mass-balance model.

FIELD OBSERVATIONS AND MODEL DESCRIPTION

Field observations

Water samples for precipitation and surface waters were observed near Yakutsk from April to August, 2000 (Fig. 1). The "Spasskaya Pad" experimental forest of the Institute for Biological Problems of the Cryolithozone (62.15°N, 129.37°E) and the "Ulakhan Sykkhan" alas (62.09°N, 130.31°E) is referred to as "Spasskaya Station" and "Alas Station", respectively. Precipitation was sampled with a funnel (diameter = 21 cm) and plastic bottle (1L). A layer of paraffin oil was put into the sample bottle so that the floating oil layer prevented evaporation of the water sample. At Spasskaya and Alas stations, precipitation was sampled every morning or soon after precipitation stopped. The precipitation at 12 other stations (A–D and R1–R8) was sampled weekly. In addition, surface waters from eight alas lakes (W1–W6, W8, and W9), the Lena River, and the Suola River (W7) were sampled once or twice per month. Lake water was sampled near the shore. Snow cover at Alas Station was sampled in April, and soil water from the surface to 90 cm depth at Spasskaya Station was sampled every month. The method of soil water sampling is described by Sugimoto *et al*. (2001). Oxygen-18 and deuterium were analyzed for over 200 of the water samples including precipitation, snow cover, soil water, river, and alas lake samples. The electrical conductivity (EC) of lake water at Alas Station was measured almost daily from May to August, and both the EC and pH of the eight other alas lakes were measured once a month in July and August. In addition, the EC and pH of precipitation and throughfall at Alas Station were measured in June and July.

To provide an isotope mass-balance analysis of a typical alas lake, intensive observations were carried out at Alas Station. This alas is situated about 50 km northeast of Yakutsk on the right bank of the Lena River; it is roughly elliptical, 1.2 km long from west to east and 0.6 km long from north to south, with a total area of 0.64 km^2. A lake with an area of less than 0.1 km^2 and no surface outflow is at the center of the alas (Ishii, 2001). Precipitation, water level of the alas lake, air temperature and relative humidity at a height of 2 m above the water surface were measured from May to September. The evaporation rate from the lake surface was calculated using the Bowen ratio method (Ishii and Yabuki, 2001). The surface area, depth and volume of the lake were surveyed once a month from May to October (Desyatkin and Desyatkin, 2001).

Isotope analysis and model description

Oxygen-18 and deuterium in the water samples were measured at the Ecological Research Center of Kyoto University, using a MAT252 with a $CO_2/H_2/H_2O$ equilibration device (Thermoquest, USA, manufactured in Germany). Isotopic concentrations are expressed as the difference between the measured ratios of the sample and reference over the measured ratio of Standard Mean Ocean Water (SMOW), expressed using delta (δ) notation:

$$\delta^{18}\mathrm{O}_{\mathrm{sample}} = \left(\frac{\left({}^{18}\mathrm{O}/{}^{16}\mathrm{O}\right)_{\mathrm{sample}}}{\left({}^{18}\mathrm{O}/{}^{16}\mathrm{O}\right)_{\mathrm{SMOW}}} - 1 \right) \times 1000(‰). \quad (1)$$

Analytical errors for the entire procedure were ±0.1‰ and 1.5‰ for δ^{18}O and δD, respectively.

According to Gibson (2001), the respective water-mass and isotope-mass balances for a well-mixed reservoir, assuming a constant water density, can be written as

$$\frac{dV}{dt} = A \times (P - E) + I \quad (2)$$

$$\frac{d(V\delta_L)}{dt} = A \times (P\delta_P - E\delta_E) + I\delta_I \quad (3)$$

where, V is the volume of the alas lake, t is time (t = 1 day), dV is the change in volume over time (dt), P is daily precipitation, E is daily evaporation, I is daily inflow, and δ_L, δ_P, δ_E, and δ_I are the δ^{18}O of the alas lake, precipitation, evaporation, and inflow, respectively. In Eqs. (2) and (3), the water-balance components (P and E) are expressed in terms of lake surface area (A), and isotopic components (δ_P and δ_L) were obtained in the field. The lack of evaporation data was accommodated by interpolation using the mean for the entire period. Temporal variation of the lake's volume was determined from changes in its water level, and the surface area was estimated from the linear function relating lake volume (data from Desyatkin and Desyatkin, 2001) and water level over each time step. There are no observations for the δ^{18}O of evaporating vapor (δ_E) at this site, because it is difficult to measure directly. In this study, δ_E was estimated as a function of dV lost by evaporation, assuming a Rayleigh process with an equilibrium fractionation factor (α*). Majoube (1971) obtained the equilibrium fractionation factor in terms of

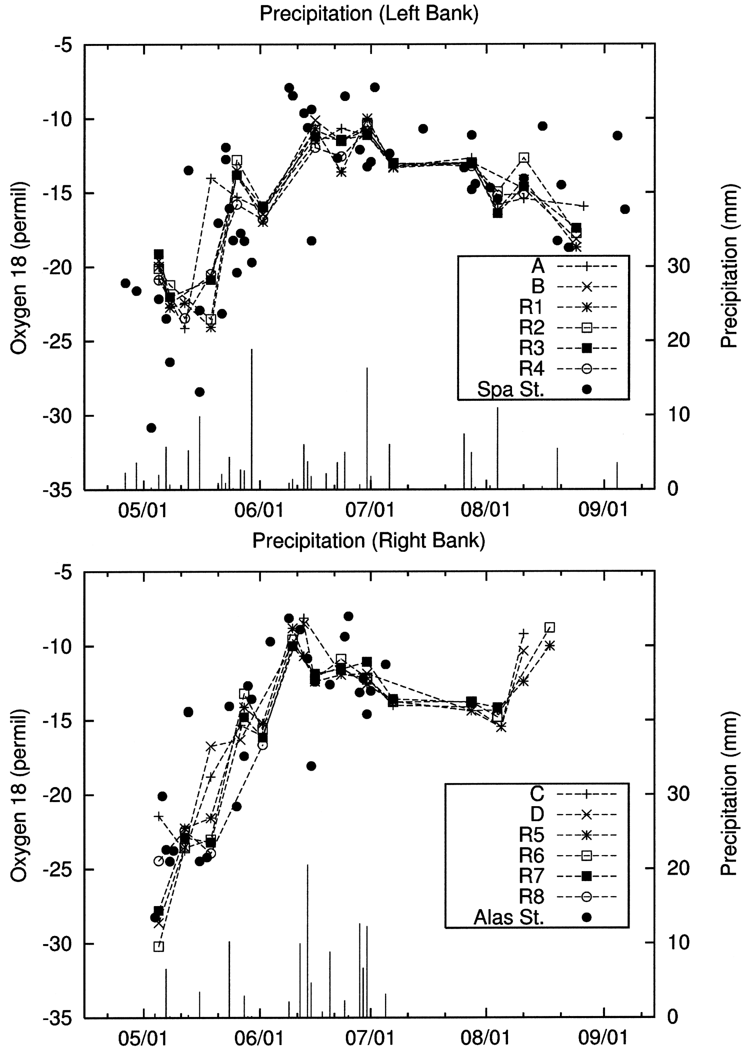

Fig. 2. Temporal variation in δ^{18}O for precipitation at Spasskaya, Alas, and the other 12 stations on the left (upper) and right (lower) banks of the Lena River. Bar graphs show daily precipitation at Spasskaya and Alas stations. Original data are from Kurita et al. (2002).

temperature (T).

$$\delta_{E,i} - \delta_{E,i-1} = 10^3\left(\alpha_i^* - 1\right) \times \ln\left(\frac{A_{i-1} \times E_i}{V_{i-1}}\right) \quad (4)$$

$$\alpha_i^* = 1.137 \times \frac{10^6}{\overline{\overline{T}}_i^2} - 0.4156 \times \frac{10^3}{\overline{\overline{T}}_i} - 2.0067. \quad (5)$$

In solving Eqs. (2)–(5), the only unknowns are I and δ_I.

RESULTS

Seasonal variation in precipitation

The temporal variation in δ^{18}O for precipitation at Spasskaya, Alas, and the other 12 stations (A–D, R1–R8; classified into the right and left banks of the Lena River), is shown in Fig. 2. The minimum δ^{18}O value for precipitation at each station ranged from approximately –30 to –20‰ at the beginning of May. The δ^{18}O value of snowfall is much lower than that of rainfall, large differences in minimum values among them were shown

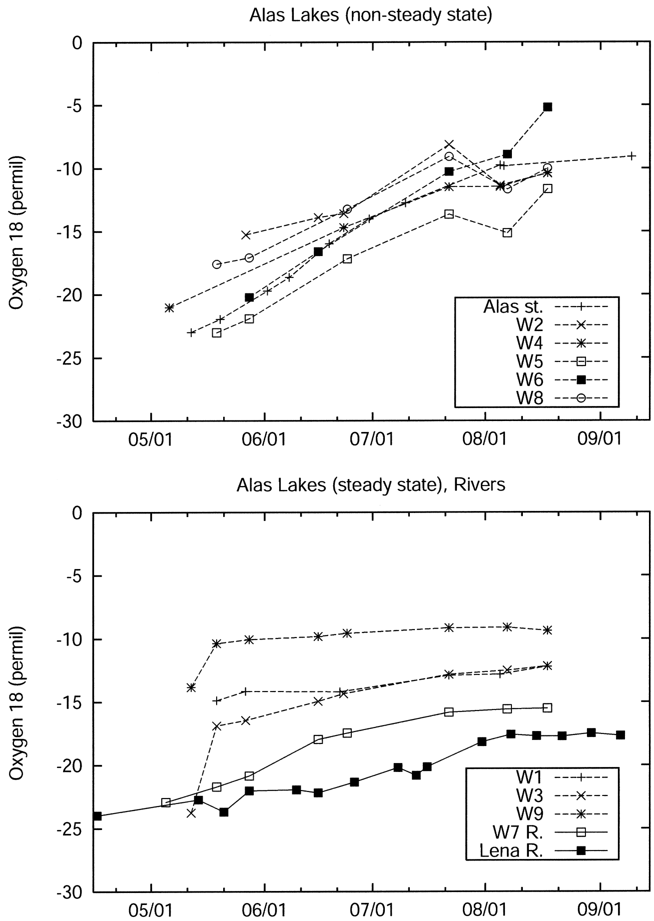

Fig. 3. Seasonal changes in oxygen isotopic compositions of non-steady-state (upper) and steady-state (lower) lakes. Values for the Lena and Suola (W7) rivers plot below the other locations.

at early spring. The δ^{18}O then rapidly increased to maximum values (approximately –10‰) by June. Values then gradually decreased to about –20‰ by August. Spatial differences in the values were very small throughout the study period following snowmelt. Kurita *et al.* (2002) pointed out that the air-mass in which rain forms is isotopically homogeneous in this region in summer, and that isotopic variation is predominantly a function of the precipitation sources.

Seasonal variation in surface water

The temporal variations of δ^{18}O in nine alas lakes and two rivers are shown in Fig. 3, and monthly averaged δD and δ^{18}O measurements are shown in Table 1. Seasonal variation of δ^{18}O in alas lakes is clearly classified into two groups. In one group, variation is of a non-steady-state type (W2, W4, W5, W6, W8, and Alas Station). The minimum δ^{18}O for these lakes ranged from –25 to –15‰ in May, whereas the maximum δ^{18}O ranged from –13 to –5‰ in July or August. The difference among the alas lakes at each observation periods exceeded 10‰. The light isotopic form of water has a higher saturation vapor pressure and diffusivity than the heavy form, and is preferentially removed from a lake via evaporation, enriching the lake water with the heavy isotope species.

In the second group, variation in δ^{18}O was

Table 1. Summary of stations and monthly averaged stable isotope compositions in surface water for 2000

Station lake or river	Alas lakes diameter stability	May δD (‰) δ^{18}O (‰)	June δD (‰) δ^{18}O (‰)	July δD (‰) δ^{18}O (‰)	August δD (‰) δ^{18}O (‰)
Alas	<100 m	–181.06	–145.97	–125.80	–109.21
lake	non-steady	–22.47	–16.31	–12.75	–9.63
W1	>1 km	–132.70	–129.19	–124.21	–120.04
lake	steady	–14.52	–14.19	–12.30	–12.19
W2	>1 km	–137.07	–128.80	–101.91	–111.78
lake	non-steady	–15.27	–18.74	–8.12	–10.40
W3	<100 m	–163.15	–139.34	–129.62	–126.77
lake	steady	–19.03	–14.65	–12.82	–12.33
W4	<100 m	–173.39	–139.32	–123.72	–116.97
lake	non-steady	–21.02	–14.70	–11.47	–10.37
W5	<1 km	–180.24	–151.55	–132.34	–124.84
lake	non-steady	–22.46	–17.17	–13.65	–13.39
W6	<1 km	–170.58	–149.18	–115.38	–100.55
lake	non-steady	–20.21	–16.60	–10.27	–7.04
W7		–168.52	–146.20	–132.86	–134.20
river	steady	–21.26	–17.71	–15.83	–15.54
W8	>1 km	–150.71	–125.37	–108.13	–115.84
lake	non-steady	–17.34	–13.23	–9.07	–10.83
W9	>1 km	–115.60	–105.33	–102.44	–103.99
lake	steady	–11.42	–9.69	–9.13	–9.22
Lena		–163.58	–156.64	–149.80	–133.16
river	steady	–21.32	–19.92	–19.48	–17.62

minimal (W1, W3 and W9). The δ^{18}O values were comparatively low (–24‰ for W3 and –14‰ for W9) during snowmelt in May, and were almost constant (*ca*. –14‰ for W1 and W3, and –10‰ for W9) after snowmelt from June to August, and the difference between the maximum and minimum became less than 5‰. The δ^{18}O for W3 was –24‰ on May 12, but suddenly enriched to –17‰ on May 19. This sudden change cannot be explained by evaporation from the lake surface. The δ^{18}O values of steady-state lakes in June were similar to the maxima for non-steady-state lakes in July or August, and close to the maximum for soil water in the lower part of the active layer (below 90 cm from the surface; Sugimoto *et al*., 2003).

Walker and Krabbenhoft (1998) examined the seasonal change in stable isotopic composition of several lakes in northern Canada. Isotopically light values were observed in early spring, whereas heavier values were observed in late fall. Substantial seasonal variation in δ values may develop in relatively shallow lakes, as compared to deeper lakes (Gonfiantini, 1986). Gibson (2001) also pointed out that the isotopic composition of large lakes was relatively stable over time. In this study, the area/depth survey was conducted only at Alas Station. In the context of the geography around the alas lakes in the study area, W1, W2, W8, and W9 are comparatively large and deep. No clear relationship was evident between lake diameter and isotopic stability (Table 1). The δ^{18}O of the Lena and Suola (W7) rivers increased slowly from –25‰ in May to –15‰ in August. The δ^{18}O values of these rivers were much lower than those of the alas lakes throughout the study period.

EC and pH variations

The EC and pH of the alas lakes and river W7 for July-August are shown in Table 2, and those of rainfall and throughfall at Alas Station for June-July are shown in Table 3. Figure 4 shows the

Table 2. The electrical conductivity (EC) and pH of alas lakes and river W7 for July and August, 2000

Station	Date	pH	EC (mS/cm)	Date	pH	EC (mS/cm)
W1	22 Jul.	8.9	0.31	5 Aug.	9.0	0.38
W2	22 Jul.	8.0	0.60	5 Aug.	8.0	0.40
W3	22 Jul.	8.0	0.80	7 Aug.	8.1	0.90
W4	22 Jul.	8.3	0.70	5 Aug.	8.1	0.70
W5	22 Jul.	7.7	0.60	7 Aug.	7.7	0.60
W6	22 Jul.	8.8	0.30	7 Aug.	7.7	0.30
W7	22 Jul.	8.3	0.40	7 Aug.	8.1	0.40
W8	22 Jul.	8.9	2.10	7 Aug.	9.2	1.80
W9	22 Jul.	9.0	0.80	7 Aug.	9.0	0.90

Table 3. The electrical conductivity (EC) and pH of precipitation and throughfall at Alas Station for June and July, 2000

Date	pH	EC (μS/cm)	Date	pH	EC (μS/cm)
Precipitation			Throughfall		
15 Jun.	4.9	14.0	15 Jun.	6.2	13.0
			20 Jun.	5.4	78.0
21 Jun.	5.4	10.0	21 Jun.	5.9	40.0
25 Jun.	6.2	13.0	25 Jun.	5.5	52.0
			26 Jun.	6.0	17.0
29 Jun.	5.9	12.0	29 Jun.	5.9	8.0
30 Jun.	6.2	8.0	30 Jun.	6.4	17.0
1 Jul.	5.0	12.0	1 Jul.	5.7	9.0
			2 Jul.	6.2	19.0
			6 Jul.	4.6	27.0

change in EC of alas lakes and the Suola River (W7) for July-August. The pH of rainfall and throughfall was less than 6.5, and that of rainfall on 15 June and throughfall on 6 July was less than 5.0. On the other hand, the pH of most alas lakes exceeded 8.0.

The maximum EC of rainfall and throughfall was below 14.0 μS/cm and 78.0 μS/cm, respectively, whereas the minimum EC of alas lakes exceeded 0.3 mS/cm (300 μS/cm). The large difference in EC among precipitation and alas lakes was observed. In August, the EC of W1–W7 and W9 was below 1.0 mS/cm, whereas that of W8 and the alas lake at Alas Station exceeded 1.5 mS/cm. The EC in the alas lake at Alas Station increased from 0.8 mS/cm in May to 1.8 mS/cm in August. This increase in EC values might resulted from evaporation from the lake surface. There was no correlation between EC and lake diameter or isotopic stability (see Table 1).

DISCUSSION

Origin of surface water

Relationships between δD and δ^{18}O in snow, river, and alas lakes are shown in Fig. 5. The δD and δ^{18}O of precipitation ranged from –30‰ for snow in April to –10‰ for summer rain in August (Fig. 2). From these data, the local meteoric water line (L.M.W.L.) obtained by the least squares method was $\delta D = 7.3 \times \delta^{18}O - 7.3$; this slope is close to that of the Global Meteoric Water Line (8.0). The local evaporation line (L.E.L.) obtained by the least squares method for non-steady-state alas lakes was $\delta D = 5.6 \times \delta^{18}O - 54.9$. This relationship is very similar to $\delta D = 5.4 \times \delta^{18}O - 55.1$,

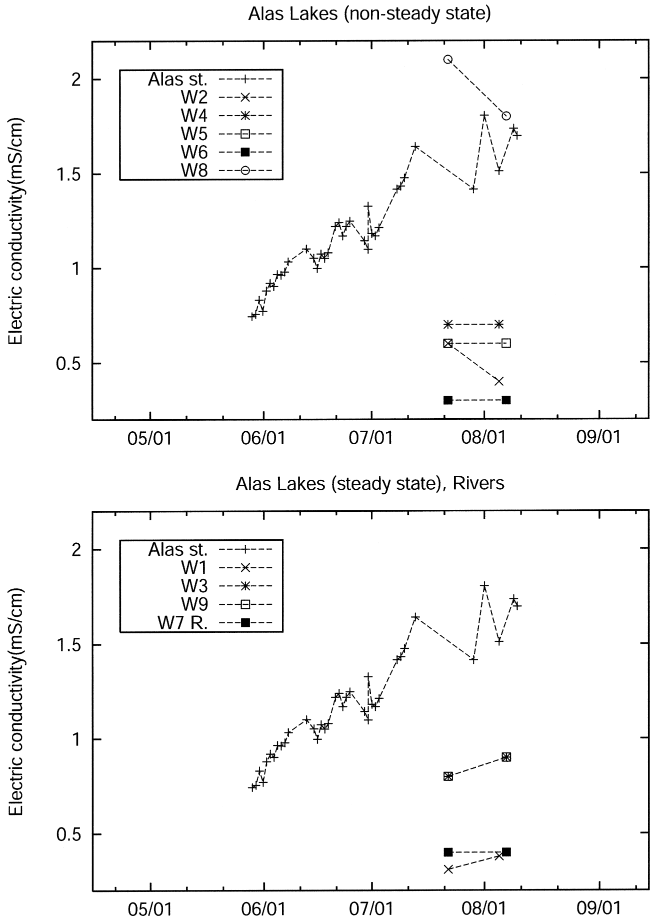

Fig. 4. Seasonal changes in the electrical conductivity (EC) of non-steady-state (upper) and steady-state (lower) lakes. Values for Alas Station and Suola River (W7) plot below the other locations.

which is the local evaporation line for surface waters in North America obtained by Gibson (2001).

Depending on the seasonal variation of δ^{18}O in summer rain, the δ^{18}O of non-steady-state lakes increased from –23‰ in May to –10‰ in August (see Fig. 3). The δD-δ^{18}O for soil water plots near the local meteoric water line. On the other hand, the δ^{18}O of steady-state lake was relatively constant and seasonal variation is minimal. Most δD-δ^{18}O plots of steady-state lakes deviated on the upper part of the local evaporation line for non-steady-state lakes, indicating that evaporation affected water in all alas lakes. Because the δD-δ^{18}O plots of non-steady state lakes in May deviated on the lower part of the local evaporation line (close to the plots for snow), the lake water likely originates from snowmelt water. After the snowmelt season, the δ^{18}O of these lakes gradually increased as a result of evaporation during the summer. Isotopic enrichment of steady-state lakes might be a result of evaporation during the previous summer. Sugimoto *et al.* (2002) measured the δ^{18}O of soil water and sap flow in 1998-2000, and pointed out that excess water present in the soil just before freezing is retained as ice during winter, and becomes available for plants the following summer or later.

The vertical mean soil water δ^{18}O weighted by the water content from the surface to 90 cm depth

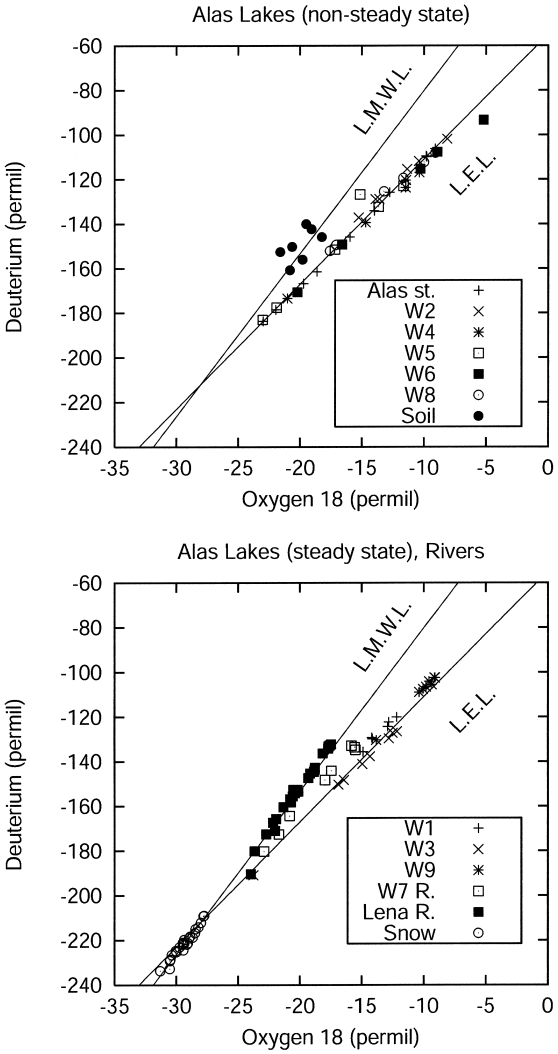

Fig. 5. Plots of δD and $\delta^{18}O$ for precipitation and surface water. The local meteoric water line (L.M.W.L.) and the local evaporation line (L.E.L.) are also shown.

plots on the local meteoric line (Fig. 5), and ranged from –23‰ in the lower part of the active layer to –18‰ in the surface layer. Although the δD-$\delta^{18}O$ of surface soil water showed evaporation effects (Sugimoto *et al.*, 2003), it does not have a significant effect on the weighted mean because the water content of surface soil is usually low.

The isotopic compositions of soil water and the Lena River lie along the local meteoric water line at Yakutsk. The $\delta^{18}O$ of water in the Lena River is thought to have the same origin as the precipitation and soil water at Yakutsk. In contrast, the isotopic compositions of the Suola River (W7) lie between the local meteoric water line and the local evaporation line of alas lakes. The Suola River water might formed by mixing of soil water and some other water that is enriched by evaporation, as in alas lakes.

Isotope mass-balance model for alas lakes

The temporal variation in the volume and isotopic composition of the lake at Alas Station can be compared with the results of the isotope mass-balance model using the Rayleigh process. Model 1 and Model 2 (Fig. 6) show the calculated results of the mass-balance model without and with an inflow component. Seasonal variation in the observed precipitation and evaporation, and the $\delta^{18}O$ of precipitation are shown in the upper panel of Fig. 6.

Observed and estimated lake volumes gradually declined from about 70,000 m^3 in May to 40,000 m^3 in August, but experienced temporary increases after heavy rainfalls. The temporal variation in lake volume estimated by the mass-balance model without an inflow component (Model 1) was lower than observed values throughout the study period. Because the volumetric difference between the Model 1 calculation and the field observation always increased by 200 m^3 every day, the Model 2 assumes a constant inflow of 200 m^3/day. Soil hydraulic conductivity ranged from 10^{-2} to 10^{-3} m/sec in the active layer (from the surface to 80 cm depth) at Alas Station (Mizoguchi *et al.*, 2001). Applying this hydraulic conductivity to Darcy's law, the assumed inflow component appears to be appropriate.

The temporal variation in the $\delta^{18}O$ of lake water estimated using the model without an inflow component (Model 1) was higher than values observed throughout the study period. This result implies that the $\delta^{18}O$ of the inflow component was lower than that of lake water. Possible sources of water in the inflow component are rainfall and soil water (subsurface flow). In the models, the precipitation contributing water to lakes

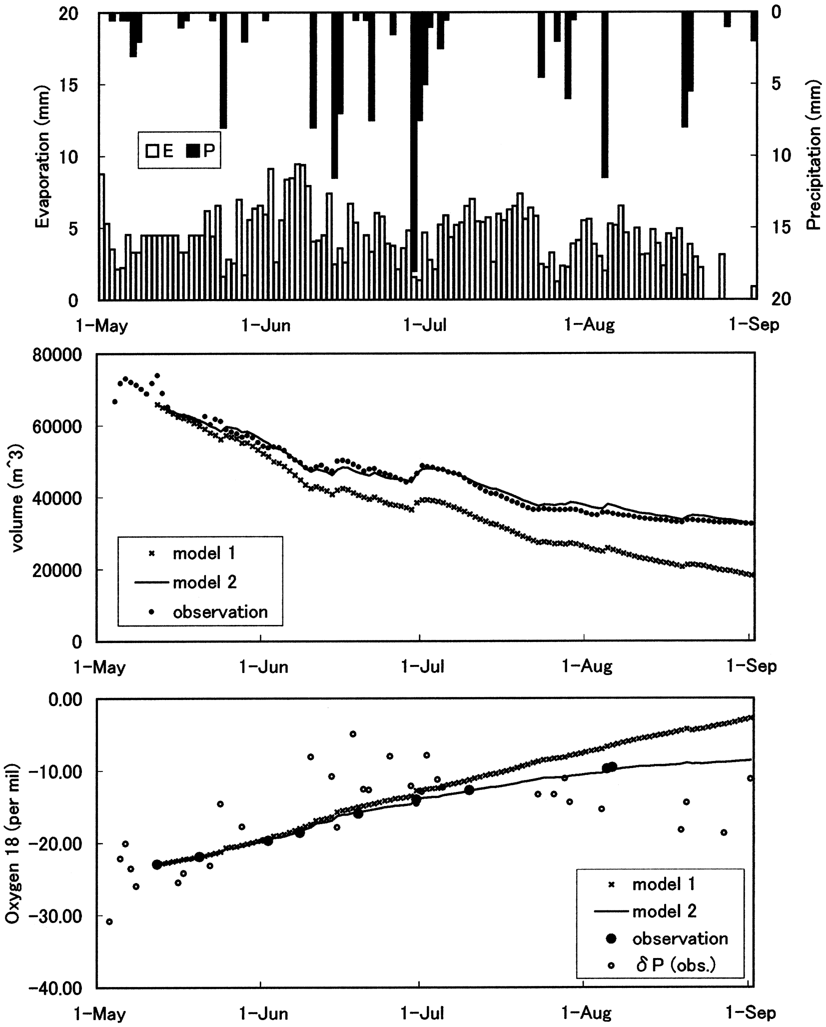

Fig. 6. Seasonal variation in precipitation (P), evaporation (E), volume, and isotopic composition of the alas lake at Alas Station. Precipitation was observed and evaporation rate was calculated using the Bowen ratio method. Simulated and observed lake volumes are shown in the middle panel. Simulated and observed $\delta^{18}O$ for the alas lake and observed precipitation are shown in the lower panel. Models 1 and 2 (middle and lower) are mass-balance models based on the Rayleigh process without and with an inflow component, respectively.

was treated as the precipitation component in Eq. (2), and calculations considered only the rainfall onto the lake surface proper. Realistically, however, the inflow component probably encompasses both precipitation onto the lake surface and that onto the surrounding alas. Rainfall δ^{18}O was not always lower than that of lake water. The actual observed δ^{18}O in rainfall from mid-June to the beginning of July was much higher than that of lake water. If rainfall with higher δ^{18}O is assumed to inflow component, calculated δ^{18}O of lake water does not gain access to the observation.

The other possible inflow component is soil water (subsurface flow). The δ^{18}O of soil water ranged from –23‰ in the lower part of the active layer to –18‰ in the surface layer (Fig. 5). The soil water content in the lower part of the active layer (below 90 cm from the surface) was much larger than that of the surface layer (Sugimoto *et al.*, 2003). Therefore, the δ_I component was assumed to be the soil water in the lower part of the active layer which has a δ^{18}O of –23‰. Substituting the inflow (I) and δ_I components into Eq. (3), the theoretical temporal variation in lake volume and δ^{18}O in the alas lake (Model 2) agreed with the observed values (Fig. 6). Substitution the δ^{18}O of soil water in the surface layer (–18‰) into δ_I component, the temporal variation of δ^{18}O in the alas lake is far from the observed values. Given that large volumes of groundwater were present in the spring of 2000 (Sugimoto *et al.*, 2003), the assumption of soil water inflow (subsurface flow) from the lower part of the active layer appears reasonable.

CONCLUSIONS

In the permafrost region of eastern Siberia, stable isotopes in precipitation and surface water were observed from May to September, 2000. The δ^{18}O variation in alas lakes were classified into two lake types: steady-state lakes in which seasonal δ^{18}O variation was less than 5‰, and non-steady-state lakes in which it exceeded 10‰. The lake water in non-steady-state lakes likely originates from snowmelt, and gradually becomes enriched in ^{18}O by evaporation during summer. In steady-state lakes, enriched lake water by evaporation in the previous summer is dominant. The local meteoric water line in this region obtained by the least squares method for precipitation was $\delta D = 7.3 \times \delta^{18}O - 7.3$, which has a slope close to that of the Global Meteoric Water Line. The local evaporation line obtained by the least squares method for alas lakes was $\delta D = 5.6 \times \delta^{18}O - 54.9$, which is very similar to that obtained at Canadian lakes (Gibson, 2001).

The volumetric and isotopic variation of alas lakes were closely reproduced by the isotope mass-balance model using Rayleigh fractionation over daily time steps. The inflow of alas lakes was estimated to be 200 m^3/day throughout the study period, and the δ^{18}O of the inflow component was assumed to be –23‰, representing soil water in the lower part of the active layer.

Acknowledgments—This research was supported by the project research in FY2000 of the Frontier Observational Research System for Global Change. The authors thank Dr. R. V. Desyatkin, Mr. A. R. Desyatkin, and Dr. A. Maximov of the Institute of Biological Problems of the Cryolithzone, Dr. R. Suzuki of the Frontier Research System for Global Change, Mr. D. Naito, a graduate student at Kyoto University, Mr. H. Tanaka, a graduate student at Tsukuba University, all members of the Alas team and GAME-Siberia for field assistance, and Ms. M. Akiho of Kyoto University for technical assistance. We also thank referees for useful comments.

REFERENCES

Desyatkin, R. V. and Desyatkin, A. R. (2001) Dynamics of alas hydrology in central Yakutia, as an index of climate fluctuations. *Proc. of GAME-Siberia Workshop, GAME Publication* **30**, 37–60.

French, H. M. (1984) *The Periglacial Environment* (translated into Japanese by Ono, Y.). Kokon-Shoin, 411 pp. (originally, 1976).

Gibson, J. J. (2001) Forest-tundra water balance signals traced by isotopic enrichment in lakes. *J. Hydrol.* **251**, 1–13.

Gibson, J. J. (2002) Short-term evaporation and water budget comparisons in shallow arctic lakes using non-steady isotope mass balance. *J. Hydrol.* **264**, 247–266.

Gonfiantini, R. (1986) Environmental isotopes in lake studies. *Handbook of Environmental Isotope Geochemistry* (Fritz, P. and Fontes, J. C., eds.), 113–168, Elsevier Science.

Ichiyanagi, K., Numaguti, A. and Kato, K. (2002) Interannual variation of stable isotopes in Antarctic precipitation in response to El Niño-Southern Oscillation. *Geophys. Res. Lett.* **29**(1), 10.1029/2000GL012815.

Ishii, Y. (2001) Outline of the field observations on the right bank of the Lena River. *Activity Report of GAME-Siberia 2000, GAME Publication* **26**, 83–86.

Ishii, Y. and Yabuki, H. (2001) Water and energy flux observation over an alas lake. *Activity Report of GAME-Siberia 2000, GAME Publication* **26**, 101–104.

Kurita, N., Numaguti, A., Sugimoto, A., Ichiyanagi, K. and Yoshida, N. (2002) The relationship between the variation of isotopic ratios and the source of summer precipitation in eastern Siberia. *J. Geophys. Res.* (accepted).

Majoube, M. (1971) Fractionnement en oxygène-18 et en deutèrium entre l'eau et sa vapeur. *J. Chim. Phys.*

197, 1426–1436.

Mizoguchi, M., Kondo, N., Tanaka, H. and Kiyosawa, H. (2001) Soil physical properties in soil profiles of active layer in Alas, Eastern Siberia. *Activity Report of GAME-Siberia 2000, GAME Publication* **26**, 111–122.

Sugimoto, A., Numaguti, A., Kurita, N., Ichiyanagi, K., Takata, K., Yamazaki, T., Suzuki, R. and Hiyama, T. (2001) Regional-scale observation on radiation, precipitation, isotopic compositions of precipitation and alas and lake water, and surface soil moisture. *Activity Report of GAME-Siberia 2000, GAME Publication* **26**, 75–82.

Sugimoto, A., Yanagisawa, N., Naito, D., Fujita, N. and Maximov, T. (2002) Importance of permafrost as a source of water for plants in east Siberian taiga. *Ecological Research* **17**, 493–503.

Sugimoto, A., Naito, D., Yanagisawa, N., Ichiyanagi, K., Kurita, N., Kubota, J., Kotake, T., Ohata, T., Maximov, T. and Fedorov, V. (2003) Characteristics of soil moisture in permafrost observed in east Siberian taiga with stable isotopes of water. *Hydrol. Proc.* **17**, 1073–1092.

Suzuki, R., Hiyama, T., Strunin, M., Ohata, T. and Koike, T. (2001) Airborne observation of land surface by video camera and spectrometers around Yakutsk. *Activity Report of GAME-Siberia 2000, GAME Publication* **26**, 61–64.

Walker, J. F. and Krabbenhoft, D. P. (1998) Groundwater and surface-water interactions in riparian and lake-dominated systems. *Isotope Tracers in Catchment Hydrology* (Kendall, C. and McDonnell, J. J., eds.), 467–488, Elsevier Science.