Chapter 6
Stable Isotopes of Water in Permafrost Ecosystem

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6.1 Introduction

6.1.1 Moisture in Permafrost Ecosystem

Climate in eastern Siberia is continental and severely dry with a mean annual precipitation of only about 230 mm. Besides, summer is not cool rather very hot, and monthly mean air temperature in July is more than 20 °C, while winter is extremely cold with a January mean air temperature of −40 °C, although winter air temperature increases recently. Under such severely dry climate, taiga that consists of deciduous conifer, larch, covers the area on permafrost (Archibold 1995).

Several species of larch are distributed in northern Eurasia, but *Larix gmelinii* and *L. cajandery* (Abaimov et al. 1998), known as Dahurian larch, form almost pure larch forest in northeastern Siberia. They are distributed not only in boreal zone but also in the Arctic zone, where soil is dry enough for them to survive (Liang et al. 2014), forming forest tundra. Southern boundary of taiga interestingly well agrees with southern boundary of permafrost distribution. This indicates that permafrost plays important role for larch forest distribution.

As explained later in this chapter, permafrost hydrology enable larch trees to survive by providing water under such dry climate as eastern Siberia. On the other hand, larch forests are essential source for atmospheric water vapor, as they transpire water. Returning of water, which fell as precipitation, from soil to the atmosphere is called "precipitation recycling" (e.g., Eltahir et al. 1998; Jasechko et al. 2013). It has been reported that atmospheric water vapor, particularly in summer in continental regions, is brought through evapotranspiration. In eastern Siberia, Numaguti (1999) showed that summer precipitation is formed by water vapor which is recycled at least twice in average by means of colored tracer technic with GCM. Hydrologic cycle in

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T. Ohta et al. (eds.), *Water-Carbon Dynamics in Eastern Siberia*,
Ecological Studies 236, https://doi.org/10.1007/978-981-13-6317-7_6

the permafrost ecosystem is figured out with the isotopic composition of water in the system.

6.1.2 Use of Stable Isotopes of Water

Stable isotopes of water have been widely used for researches on water cycle, particularly to investigate source of water, including water vapor, because water isotopes can be used as tracer of water (e.g., Gat 2000). Most of researches on precipitation recycling have applied stable isotopes of precipitation and water vapor to trace water (Jasechko et al. 2013; Kurita et al. 2003, 2004). In addition, stable isotopes of sap water in plants have been also used to know the source of water for plants in ecological studies (e.g., Dawson and Ehleringer 1991; Querejeta et al. 2007).

Applications of stable isotopes of water are generally made with isotope ratio of oxygen ($^{18}O/^{16}O$) and hydrogen (D/H) of water, and these isotope ratios are expressed with $\delta^{18}O$ and δD for oxygen and hydrogen, respectively, defined as a permil deviation from an international standard as below.

$$\delta^{18}\mathrm{O}\ \text{ or }\ \delta\mathrm{D}\ = [\mathrm{R_{SA}}/\mathrm{R_{SMOW}} - 1] \times 1000\ (‰),$$

where R_{SA} and R_{SMOW} are isotope ratios of oxygen or hydrogen for sample and international standard SMOW (standard mean ocean water). Adding to these δ values, another parameter d-excess (Dansgaard 1964) is used:

$$\text{d-excess} = \delta\mathrm{D} - 8 \cdot \delta^{18}\mathrm{O}\ (‰).$$

The d-excess has been widely used in many studies for mainly two different ways. One is to know a source of water vapor (e.g., Tian et al. 2001), because this value results from evaporation process of water vapor on sea surface and is conservative in condensation process. Other applications have been made for investigations on evaporation process from surface (e.g., Gibson et al. 2005; Gibson and Reid 2014).

It has been well known that isotopic composition of precipitation observed in most regions in the world is generally on a line, $\delta D = 8 \cdot \delta^{18}O + 10$, in $\delta D - \delta^{18}O$ plot, so-called meteoric water line. When liquid water is formed from water vapor, heavy isotopes (HDO and $H_2{}^{18}O$) condensate faster than light isotope ($H_2{}^{16}O$), because of a slightly different saturation vapor pressure among three isotopes. Therefore, as precipitation process proceeds, the δD and $\delta^{18}O$ values of water vapor gradually decrease, because heavy isotopes are removed from water vapor and enter into liquid phase, resulting in gradual decreases in δD and $\delta^{18}O$ of precipitation (liquid phase) which is formed from.

6.2 Water Budget of Taiga Forest Ecosystem

6.2.1 Changes in Soil Moisture and Its Water Isotopic Composition

6.2.1.1 Isotopic Composition of Precipitation in Eastern Siberia

Source of water in an ecosystem is basically precipitation. Interestingly, the isotopic composition of precipitation in this region shows outstanding seasonality like a seasonal variation in air temperature, with extremely low values during winter while high values in summer. Figure 6.1a shows the δ^{18}O of precipitation obtained daily sampling in 1998 and 1999. During summer (June, July, and August), precipitation shows high δ^{18}O values (mostly −15 to −10‰), which is not different from precipitation observed in mid-latitude region, whereas those in winter are extremely low (mostly lower than −30‰). This seasonality can be explained by the difference in precipitation recycling between summer and winter, as shown in Fig. 6.1b. In winter, evapotranspiration is negligible because of low temperature and no vegetation activity, resulting in continuous removal of heavy isotopes from water vapor through snowfall, leading to a remarkable decrease in δ^{18}O of snow inland, whereas

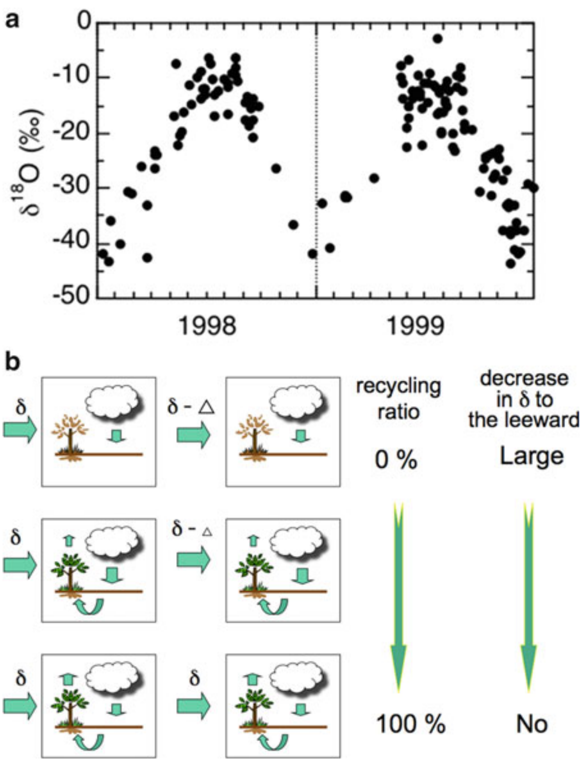

Fig. 6.1 The δ^{18}O values of precipitation sampled daily in 1998 and 1999 at Yakutsk (**a**) and schematic figure showing difference in recycling ratio between winter and summer (**b**). (The data shown in (**a**) is from Sugimoto and Maximov 2012)

in summer, high recycling ratio (large fraction of rainwater returning to the atmosphere) causes less decrease in δ^{18}O of water vapor and rainwater inland. If 100% of rainwater returned to the atmosphere through transpiration, no decrease in the δ^{18}O of water vapor would occur. As we usually observed seasonality for precipitation isotopes values, temperature difference in condensation process between summer and winter is also the secondary reason.

6.2.1.2 Soil Moisture Equivalent and the Water Isotopes

Water budget at a site can be written by a simple equation as below with a unit mm, when lateral flow is negligible:

$$P = ET + R + \Delta Qs,$$

where P, ET, and R are the amounts of precipitation, evapotranspiration, and runoff and the last term ΔQs is a change in the soil moisture equivalent. On an annual time scale, ΔQs is negligible in general. However, in a cold region especially in permafrost region, this ΔQs (annual change) is significantly large (e.g., Troy et al. 2011), and this is one of distinguishing features observed for soil moisture in this region and another cold region as well, although the long-term average of ΔQs is zero under stable climate. As seen in Fig. 6.2, soil moisture equivalent observed at Spasskaya Pad (see Sect. 3.2) shows large inter-annual variations.

In an observation of soil moisture, TDR (time-domain reflectance) sensor is frequently used. This sensor detects only liquid phase of water, and ice is not detected, since dielectric property of ice is not different from another materials in soil such as minerals. Therefore, soil moisture in a cold region, where the soil freezes in winter, can be observed only in summer, and inconveniently or conveniently TDR detects ice melt process in the soil. From a point of view of water budget, both amounts of ice and liquid water in the soil are necessary. Soil moisture equivalent shown in Fig. 6.2 includes both liquid water and ice in the soil, calculated by the same method by Sugimoto et al. (2003). They successfully calculated total soil moisture equivalent including ice, on the basis of the fact that soil moisture in the end of summer is kept as ice in the soil until the following summer. In this region, water is seldom to discharge from the site in fall and winter, because of impermeable property of frozen soil. Therefore, in most years, soil moisture equivalent in a soil layer before it freezes is almost equal to that after it thaws in the following summer, except for the surface soil layer in which snowmelt water infiltrates. Soil moisture equivalent including ice was thus estimated, assuming no change in it during freezing period if there is no difference in it between fall (before freeze) and following summer (after thaw).

Estimated soil moisture equivalent (Fig. 6.2) showed large intra- and inter-annual variations. In May in most years, surface soil layer (0–15 cm) showed clear increase in equivalent, indicating snowmelt water infiltrated into the layer. This was also supported by the observed decrease in δ^{18}O of soil water for that layer, resulted from infiltration of snowmelt water with very low δ^{18}O (Sugimoto et al. 2003). From an

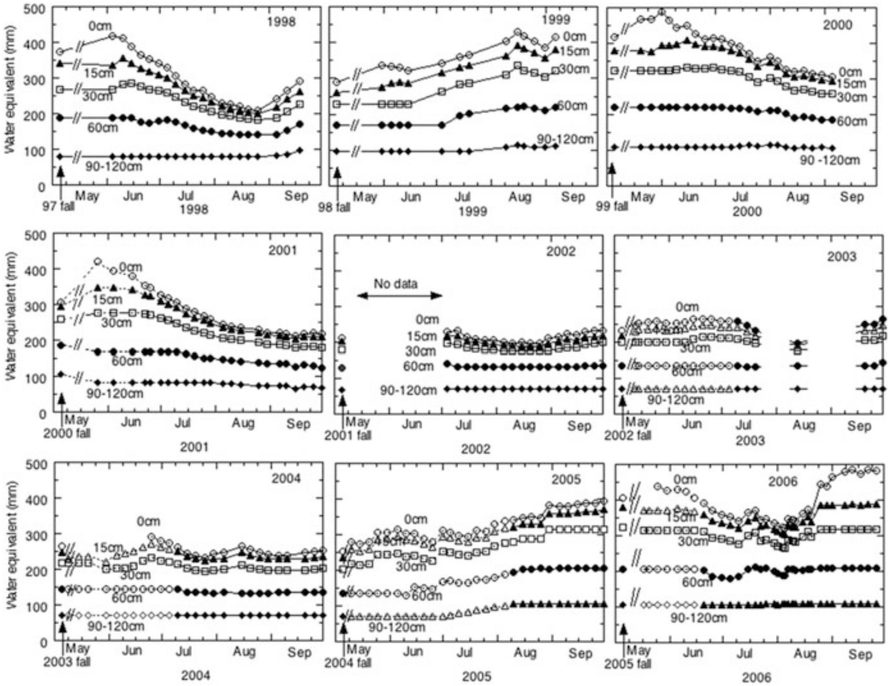

Fig. 6.2 Intra- and inter-annual variation of soil moisture water equivalent including ice from surface of mineral soil layer to 120 cm. Soil moisture water equivalent was estimated with the same method by Sugimoto et al. (2003). Data for 1998–2000 were obtained from their study

increase in the amount of snow equivalent and decrease in the δ^{18}O of soil moisture, fraction of snowmelt water which reaches to the mineral soil layer was estimated to be about 50%. The other half of snowmelt water is expected to be lost through evapotranspiration from surface organic layer and/or discharge from the site before reaching to the mineral soil layer.

After this increase in soil moisture equivalent by infiltration of snowmelt water, its variation clearly reflects summer rainfall (Fig. 6.2). In summer with large amount of rainfall (years 1999, 2005, and 2006), soil moisture increases, whereas in drought summer it decreases (years 1998, 2000, and 2001), depending on the balance between P and ET. Another important feature of soil moisture is the carry-over of soil moisture from fall to the following spring. By the end of warm season (September) in wet years, soil moisture increases, and in the following year, soil moisture variation starts from high level of soil moisture. This means that soil moisture available for vegetation depends on not only the amount of precipitation in the year but also on the amount of moisture carried over from the previous summer. As described below, the surface soil moisture equivalent differed from year to year, but in longer time scales, equivalent in the length of saw depth varied, and their isotopes also changed slightly.

As expected and also reported in many places, the $\delta^{18}O$ of soil water in shallow soil layer shows significant temporal variation depending on a fluctuation of isotopic composition of precipitation and subsequent evaporation from surface, while variability of isotopic composition of soil water decreases with depth and converges to a specific value (e.g., Hsieh et al. 1998). Figure 6.3a shows the $\delta^{18}O$ value of soil moisture including ice observed at Spasskaya Pad for the depth from 80 to 200 cm, which is the bottom part of active layer and upper part of permafrost. Obviously, the $\delta^{18}O$ of this depth of soil layer is almost constant (around −24‰). Active layer depth at this site is usually 1.2–1.4 m; therefore lower half of this figure is permafrost. The $\delta^{18}O$ values shown in this data figure (−24‰) are between summer rainfall and winter snowfall and are close to annual average value or slightly lower than the annual average, which is expected to be an average for water percolated into the deeper layer.

The $\delta^{18}O$ value of deeper soil layer converges to a specific value as seen in Fig. 6.3a, and variation of that value is not large, but may change slightly with a longer time scale. Figure 6.3c is an example of change in the $\delta^{18}O$ of soil water in deeper soil layer, showing that one of the vertical profiles of $\delta^{18}O$ observed in 2000 shifted to higher value. This is interpreted as percolation of summer rainwater fell in 1999 in sandy soil layer. The $\delta^{18}O$ of soil moisture (water and ice) in deeper layer is conservative but may slightly change depending on the water budget at the site with longer time scale. When summer precipitation exceeds evapotranspiration, it may shift to higher value, whereas it may shift to lower value when evapotranspiration exceeds precipitation.

6.2.2 *Source of Water for Plants*

Sap water that was absorbed from roots is transported to leaves in which transpiration occurs. During this transportation, no isotopic change occurs in general, though a transpiration enriches heavy isotopes in leaf water (Yakir and Sternberg 2000). This fact enables us to use sap water $\delta^{18}O$ to know source water for the plant. In addition in eastern Siberia, as described in Sect. 6.2.1, the $\delta^{18}O$ of precipitation shows extremely large seasonality with high $\delta^{18}O$ in summer rainfall and extremely low value in winter snowfall. This seasonality of $\delta^{18}O$ gives us advantages for applications with isotope ratios of water in ecosystem, by tagging water with different $\delta^{18}O$ values.

Sap water $\delta^{18}O$ of larch trees observed in June and August at Spasskaya Pad showed large intra- and inter-annual variations, ranging from about −23 to −12‰ (Fig. 6.4a). Observed $\delta^{18}O$ in June is low, and its variation is not so large compared to that in August, indicating that plant could use snow meltwater with low $\delta^{18}O$ every year (Sugimoto et al. 2002). On the other hand, large variation is found in August with very high $\delta^{18}O$ in wet summer and with low values in dry summer as described in Sect. 6.1, reflecting the $\delta^{18}O$ of soil moisture which was used by larch trees. In wet summer with enough amount of rainfall, surface soil is recharged with summer rainwater of which $\delta^{18}O$ is high, whereas in dry summer when the amount

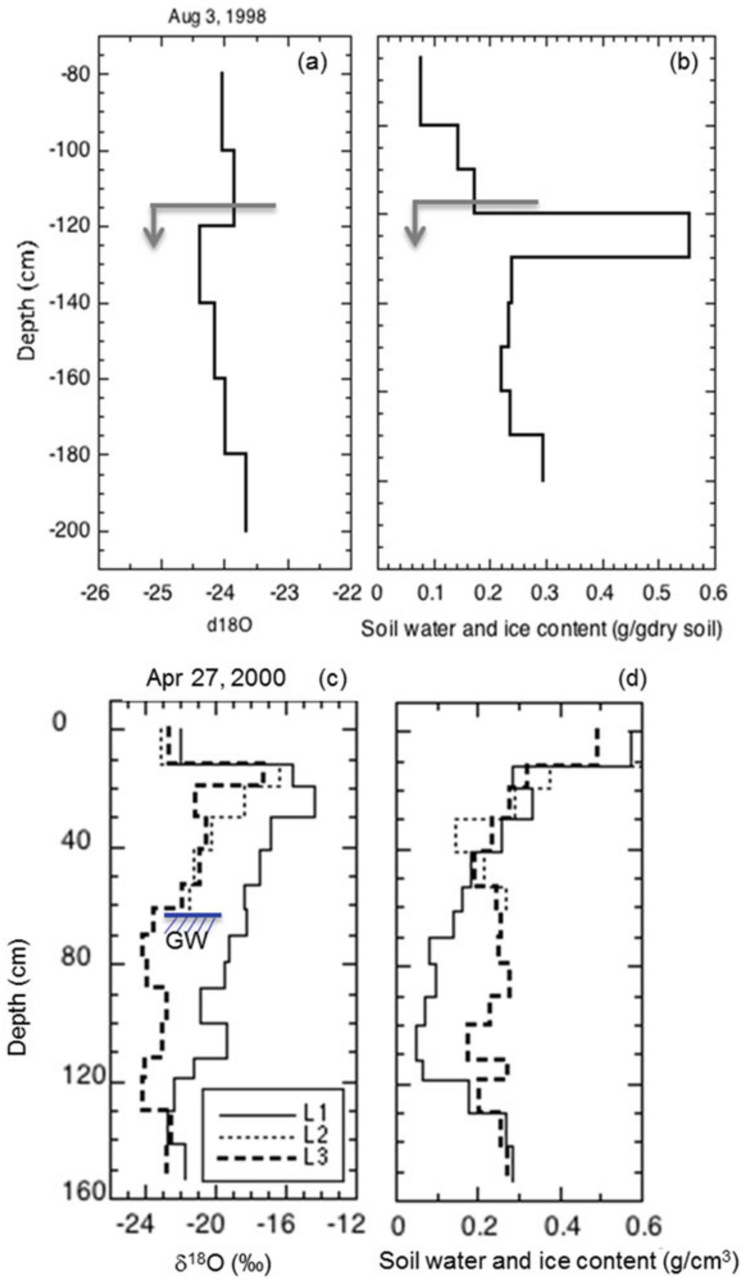

Fig. 6.3 The $\delta^{18}O$ value (**a**) and water content (**b**) of soil moisture (water and ice) in the layer from 80 to 200 cm observed on August 3 in 1998 and those from surface to 150 cm observed on April 27 in 2000 (**c** and **d**). Soil layer shown in (**a**) corresponds to lower half of active layer (80–140 cm) and upper part of permafrost (140–200 cm). GW in (**c**) is a water table of talik which was formed after heavy rainfall in 1999. (Data were modified from Sugimoto et al. 2003)

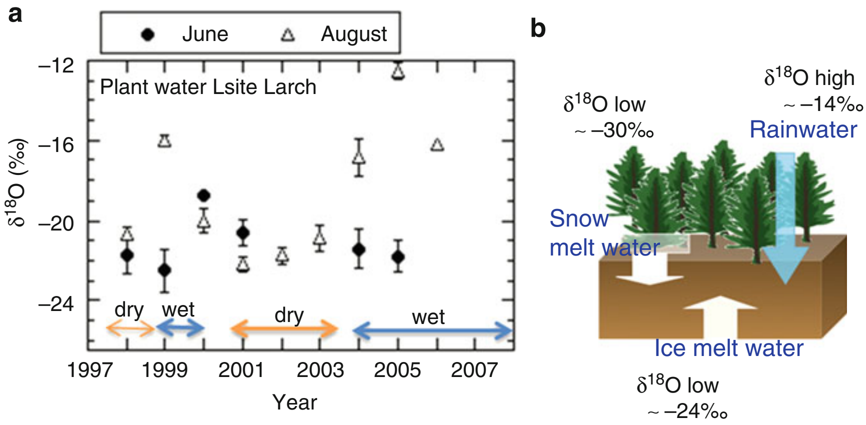

Fig. 6.4 Sap water $\delta^{18}O$ of larch trees (**a**) observed at Spasskaya Pad in June and August for years from 1998 to 2006 with description on the soil moisture condition (wet or dry) and schematic figure on source of water for larch trees (**b**), showing three sources of water for plants, snow meltwater, summer rainfall, and ice meltwater in the soil

of summer rainfall is not enough, soil water produced from ice meltwater in the soil of which $\delta^{18}O$ is low (around −24‰) is used by larch trees, resulting in the decrease in sap water $\delta^{18}O$.

This fact indicates that storage of water which was carried over from the previous fall is an important source of water for plants in drought summer, and in other words, soil moisture stored as ice during summer is an important source of water for plants during drought.

In summary, there are three sources of water for plants, snowmelt water, summer rainwater, and ice meltwater in the soil (Fig. 6.4b).

6.2.3 *Role of Water and Ice in the Bottom Layer of Active Layer and Uppermost Layer of Permafrost*

The $\delta^{18}O$ value of soil moisture (water and ice) shows large intra- and inter-annual variations in surface soil layer, while that in the deeper soil layer shows less variation, as already explained in Sect. 6.1. The same is true for the amount of soil moisture. Total amount of soil water and ice in the surface soil layer varies significantly seasonally and also inter-annually, while less variations are seen in deeper soil layer (60–120 cm) in Fig. 6.1. However, soil moisture in this layer also varies with longer time scale, depending on the water budget (Sugimoto et al. 2003), as described already.

Figure 6.5 shows the soil moisture equivalent (water and ice) and its $\delta^{18}O$ for every 30 cm depth of soil layer from surface to 150 cm depths observed in warm season, with the amount and the $\delta^{18}O$ of snowfall in previous winter and that in

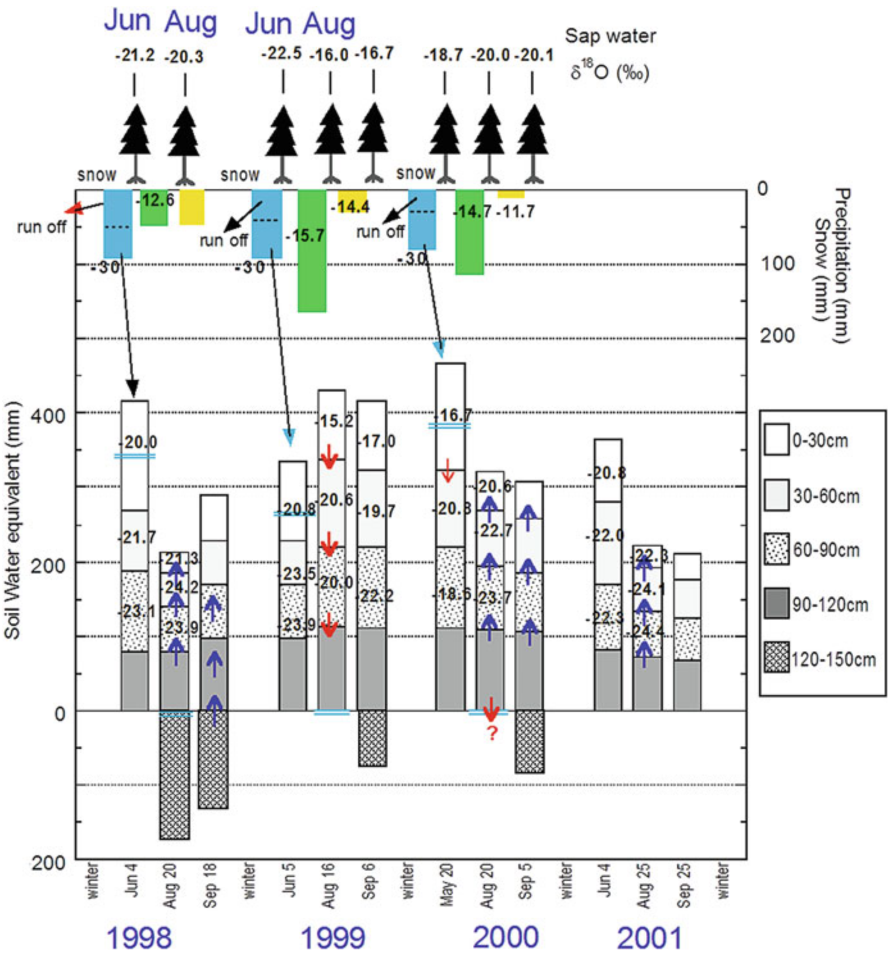

Fig. 6.5 Soil moisture water equivalent (water and ice) for every 30 cm of soil layer from surface to 150 cm with the $\delta^{18}O$ data. The amount of snowfall in previous winter and the amounts of rainfall in summer (June, July, and August) and September with their $\delta^{18}O$ are also shown. Schematic figure on the top of the figure shows the $\delta^{18}O$ of sap water of larch trees, that is, the $\delta^{18}O$ of water vapor transpired to the atmosphere. (Data for 1998–2000 were obtained from Sugimoto et al. 2003)

summer and fall. Change in the soil moisture equivalent and its $\delta^{18}O$ clearly depend on the summer precipitation. In wet summer with large amount of rainfall having high $\delta^{18}O$ in 1999, increase in the amount of soil moisture equivalent was observed in August with an increase in $\delta^{18}O$, indicating downward water transportation. On the other hand, in dry summer (1998, 2000, and 2001), decreases in soil moisture equivalent and $\delta^{18}O$ were seen, suggesting upward transportation of soil moisture.

As seen in Fig. 6.3b, soil moisture (water and ice) content in the bottom of active layer is high, and ice lenses are frequently observed. Thaw of this soil layer produces liquid water, and it can be a source of plant during drought in the following year but not in the year, because soil thaw of this deep layer occurs in fall (September and the beginning of October), in other words after plant growing season was over. When liquid water is produced by soil thaw, and if upper soil layer is dry, it may be transported upward and carried over until next summer, which makes possibility to be used by plants in the following year.

The uppermost layer of permafrost, which has also high moisture (ice and water) content, plays the same role as that in the bottom of active layer as described above, because this layer and the bottom of active layer occasionally thaw (became active layer) and occasionally remain frozen for more than 2 years (permafrost). Therefore, from a point of view of water budget, these layers are considered as water storage with longer time scale.

These layers are conceptually the same as "transition zone" defined by Shur et al. (2005) which act as buffer between active layer and permafrost and protect permafrost by providing latent heat required for thaw. These layers also act as a moisture storage with longer time scale than that of upper part of active layer.

6.2.4 Discharge of Water from Land to River

Because of a dry climate, soil moisture decreases during summer season in many years at Spasskaya Pad (Fig. 6.2), indicating that evapotranspiration exceeds preparation in the site scale. In such case no discharge from land to river is expected. In addition, discharge through groundwater seldom occurs in this region because of impermeable property of permafrost. Discharge from land to river is produced through shallow soil layer, if it occurred, and it may happen only in very wet years when soil moisture is kept unfrozen in the soil as seen in Fig. 6.3c, d. Discharge from land to alas in wet year (1999–2000) was also pointed out from a calculation of isotope mass balance model on alas water (Ichiyanagi et al. 2003).

While the water budget on site scale indicates that precipitation exceeds evapotranspiration during summer, basin scale water budget shows discharge generation during summer (Serreze et al. 2002; Oshima et al. 2015), although there are still large uncertainties in baseline data set for estimation of water budget (e.g., Troy et al. 2011; Bring et al. 2016). Basin scale water budget reflects budget in each part of the river basin. Majority of discharge of the Lena river is generated in southern mountain taiga region (Ma et al. 2000).

The isotopic composition of the Lena river water exhibits a clear seasonal variation with a pulse of low $\delta^{18}O$ generated by snowmelt, followed by increases during summer and then decreases gradually in fall and winter, and shows relatively constant value around −20‰ (Fig. 6.6a). This seasonality of $\delta^{18}O$ allows us to figure out an outline of hydrological system of the Lena river basin. Obviously

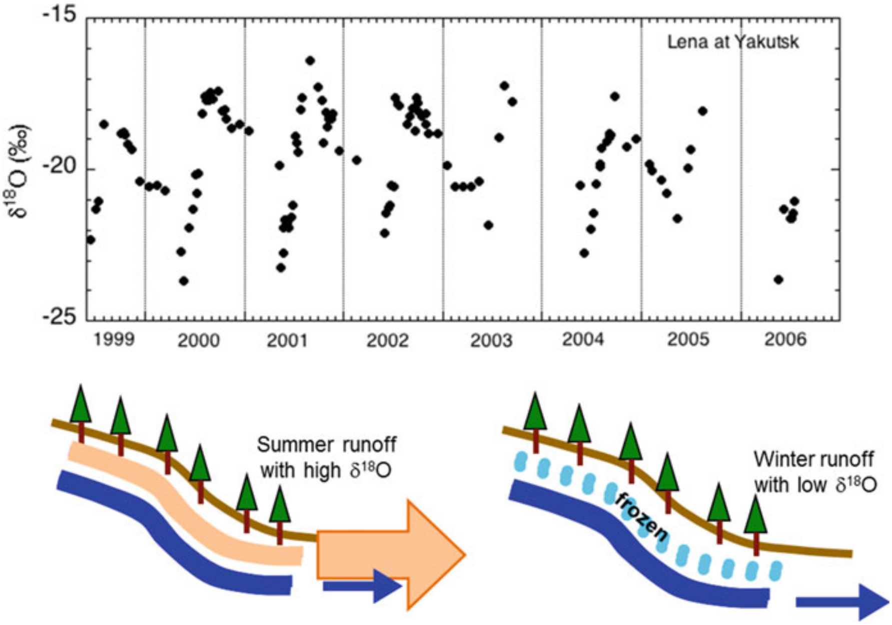

Fig. 6.6 Temporal variation of $\delta^{18}O$ of Lena river observed at Yakutsk (**a**) and schematic figures which explain runoff (**b**) in summer (left) and in winter (right). (Reproduced from Sugimoto and Maximov 2012)

summer runoff with high $\delta^{18}O$ water is generated in surface soil layer, having a large contribution of summer rainfall, whereas winter runoff with relatively low $\delta^{18}O$ seems to form unfrozen soil moisture in deeper soil layer below frozen surface soil, in southern mountain taiga region (Fig. 6.6b).

Welp et al. (2005) observed a similar seasonal variation in the isotopic composition of the Kolyma river water. They estimated fractions of snow origin and rain origin water in river water with the $\delta^{18}O$ values of those end members and revealed that substantial portion of snowmelt water remains in the basin and returns to the atmosphere through transpiration and also run off during summer. Their calculated results show that increase in the $\delta^{18}O$ of river water during summer after the pulse of snowmelt implies the decrease in fraction of snow origin water and increase in that of rain origin water.

Increase in Arctic river discharge especially in winter base flow has been reported, particularly in northeastern Eurasian rivers for which permafrost is observed in the basins (Peterson et al. 2002; McClelland et al. 2006). The isotopic composition of river water reflects of origin of runoff water as shown by Wepl et al. (2005). River isotope data may help us to figure out hydrological changes which may proceed invisibly.

6.3 Taiga as a Source of Atmospheric Water Vapor

As expected from the studies on precipitation recycling in eastern Siberia (Numaguti 1999; Kurita et al. 2003, 2004), water vapor transpired by plants contributes significantly to the atmosphere. Precipitation recycling is the good example to use water isotopes and is important terrestrial water fluxes not only for eastern Siberia but also the inland regions on the globes (Gat 2000; Jasechko et al. 2013; Wang et al. 2016; Winnick et al. 2014). Basically, there are two different usages of isotopes for the precipitation recycling. Using the d-excess and delta values of precipitation and/or atmospheric water vapor, transpiration and evaporation were separated (Jasechko et al. 2013; Wang et al. 2016), because transpiration does not change the isotope ratio for larger spatial and longer time scales, while evaporation makes water vapor with lower isotope ratios (Gat 2000). On the other hand, the isotopes of precipitation decrease inland, which is called continental effect. These continental effects showed different ways of decreasing, by adding the transpiration (Winnick et al. 2014).

Transpiration of vegetation is also important drivers of water cycle in such dry-climate region as eastern Siberia. Transpiration (precipitation recycling) has been also observed at Spasskaya Pad. Direct field observations of the isotopic composition of water vapor conducted in 2006, 2007, and 2008 at Spasskaya Pad revealed the importance of plant transpiration as a source of atmospheric water vapor (Ueta et al. 2013, 2014). Typical diurnal variation in the $\delta^{18}O$ of atmospheric water vapor on clear day observed at Spasskaya Pad was increased in the $\delta^{18}O$ in the morning with increasing mixing ratio and decreased in the afternoon (Fig. 6.7a, b). As shown in Fig. 6.7c, d, soil water with the $\delta^{18}O$ ranging from −18.8 to −11.2‰ was expected to be transpired by plants, and Ueta et al. (2013) report that the $\delta^{18}O$ of sap water of larch trees ranged from −17.9 to −13.3‰. The values for plants and soil $\delta^{18}O$ are considerably higher than those of the atmospheric water vapor that ranged from −29.7 to −18.4‰. Therefore, observed increase in the $\delta^{18}O$ values of water vapor with the increase of mixing ratio in the morning is explained by the addition of transpired water vapor with high $\delta^{18}O$ value. On the other hand, in the afternoon, decrease in $\delta^{18}O$ with decreasing mixing ratio is caused by entrainment of free atmosphere in which water vapor is expected to have a low $\delta^{18}O$.

Clear correlation between $\delta^{18}O$ of atmospheric water vapor and mixing ratio observed at midday has been found in midsummer (July) in 2006 and 2008 not only in diurnal time scale but also monthly time scale at Spasskaya Pad (Ueta et al. 2014), as an example for 2008 shown in Fig. 6.8. Clear correlation has been also observed between the $\delta^{18}O$ and d-excess. Simultaneous increases in the $\delta^{18}O$ and mixing ratio are observed when air temperature increases. This is also interpreted as a mixing of transpired vapor with high $\delta^{18}O$ and background water vapor with low $\delta^{18}O$. When air temperature increases, resulting in active transpiration, as a result, increase in contribution of transpired water vapor causes a high $\delta^{18}O$. Correlations of the $\delta^{18}O$ with mixing ratio and d-excess indicate a mixing of transpired water vapor with a background atmospheric water vapor, because the latter is expected to have a low $\delta^{18}O$ with a relatively high d-excess (Bariac et al. 1990). Assuming the $\delta^{18}O$ values

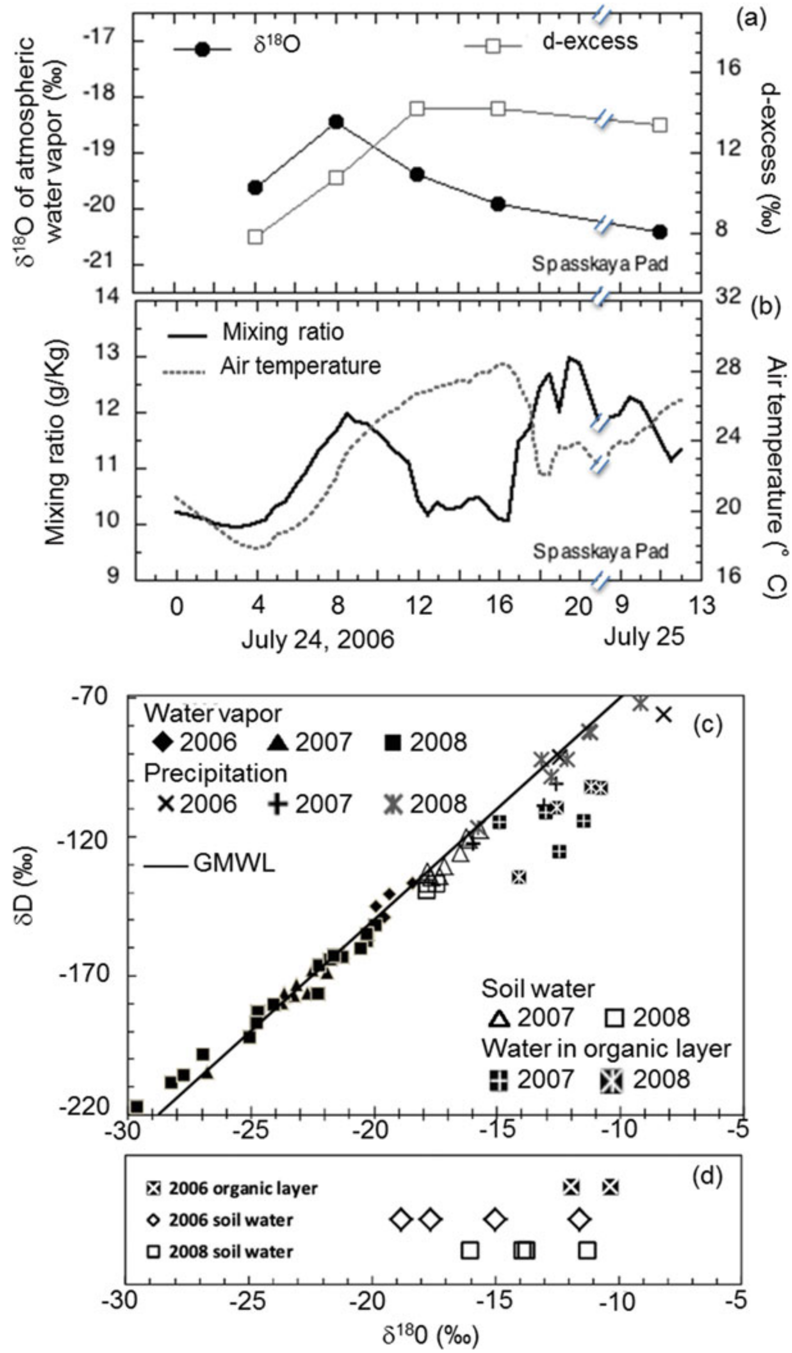

Fig. 6.7 Typical diurnal variations of $\delta^{18}O$ and d-excess (**a**) of atmospheric water vapor and air temperature and mixing ratio at 24 m (**b**) observed at Spasskaya Pad on fine day and δD- $\delta^{18}O$ plots for various water samples (**c**) and $\delta^{18}O$ of surface soil water (**d**). (Reproduced from Ueta et al. 2013)

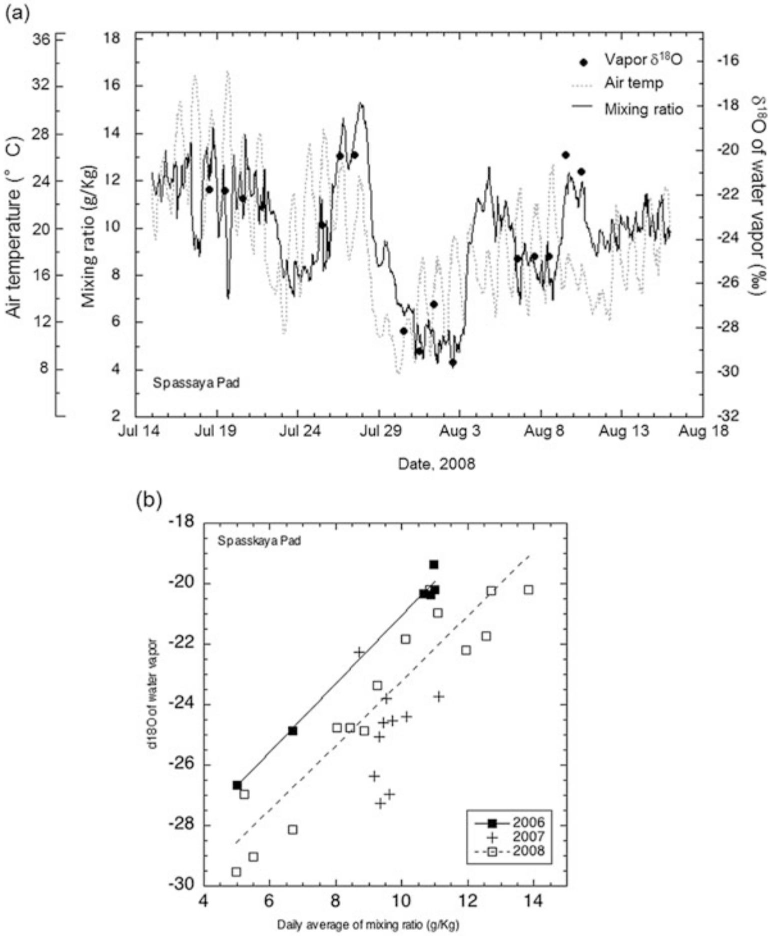

Fig. 6.8 The δ^{18}O values of atmospheric water vapor, mixing ratio, and temperature observed at canopy level in 2008 at Spasskaya Pad (**a**) and relationship between δ^{18}O and mixing ratio observed in 2006 (closed squares), 2007 (+), and 2008 (open squares). (Reproduced from Ueta et al. 2014)

of transpired vapor and background atmospheric water vapor, fraction of water vapor derived from transpiration is estimated to be up to 80%. Correlations of the δ^{18}O with mixing ratio usually occur during the summer, responding to the advection from any direction (Ueta et al. 2014). This suggests that water vapor from transpiration may be an important vapor source for precipitation.

Interestingly, correlation between δ^{18}O of atmospheric water vapor and mixing ratio was not observed in 2007 when extreme wet event occurred in this region, and waterlogging happened at depressions in the forest (Ueta et al. 2014). Sap flow rate observed in that year was very low and showed no response to precipitation. Many trees growing at depressions died in this extreme wet event, although the trees living at ridges and well-drained sites grew well with plenty of water available in the soil. In this extreme wet condition, evaporation from the surface, not transpiration, may contribute to the atmospheric water vapor.

6.4 Concluding Remarks

Permafrost plays various important roles in water cycle in eastern Siberia. Water tagged with low $\delta^{18}O$ by snowmelt is useful tracer. Uppermost layer of permafrost is also tagged with this water, and soil moisture in the bottom of active layer has also the same $\delta^{18}O$. Continuity of this $\delta^{18}O$ of soil moisture (ice and water) indicates that "transition zone" consisting of bottom of active layer and uppermost layer of permafrost should also be considered from an isotopic point of view. This zone plays a role of water storage with a longer time scale (more than 5 years), because moisture in this layer seems to provide water to the middle active layer (about 60 cm depth at Spasskaya Pad), and to that depth larch tree roots may reach. During a long drought period (2001–2003), sap water $\delta^{18}O$ of larch trees became low, and soil moisture $\delta^{18}O$ in the bottom of active layer decreased, indicating that moisture in the transition zone provides water to larch trees.

Currently, transition zone plays a role of water storage with longer time scale and provides water to the depth reachable for larch roots. However, if active layer and transition zone deepen in the future, moisture in this zone may not be available for larch trees. On the other hand, extreme wet event also threatens the forest.

As will be described in Chap. 7, soil moisture reconstructed from carbon isotope ratio of larch tree ring shows that extreme wet event observed in 2007 was the wettest event in past 100 years in this forest (Tei et al. 2013). This extreme wet event made great impacts on water and carbon cycles in this forest, and frequency of such extreme wet event as observed in 2007 may increase under warming condition in the future (Hartmann et al. 2013; Collins et al. 2013).

Stable isotopes of water figured out the water cycle of permafrost ecosystem. Vegetation (larch trees) is one of the components of permafrost hydrologic system. It obviously plays an important role in precipitation recycling, and at the same time, it is also supported by the permafrost hydrologic system.

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