



Permafrost Hydrology: Linkages and Feedbacks

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Abstract

In the cold regions, hydrological regime is closely related with permafrost conditions, such as permafrost extent and thermal characteristics. Ice-rich permafrost has a very low hydraulic conductivity and commonly acts as a barrier to deeper groundwater recharge or as a confining layer to deeper aquifers. In regions underlain by permafrost, the active layer is the upper layer of the soil near the surface that undergoes thawing in the summer and freezing in the fall. The thawing starts from the surface in the spring, and the active layer reaches its maximum in late summer. The lower boundary of this layer is the top of the permafrost layer. The active layer is considered to produce base flow (or low flow) during the ice-free season. In this chapter, we discuss relationship between permafrost coverage and streamflow regime, detection of permafrost thawing trends from long-term streamflow data, determination of permafrost groundwater age, and water balance of northern permafrost basins.

Keywords

Permafrost coverage · Base flow (low flow) · Basin (terrestrial) water storage · Permafrost thawing trends · Groundwater age · Water balance

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16.1 Introduction

In the cold regions, hydrological regime is closely related with permafrost conditions, such as permafrost extent and thermal characteristics. Ice-rich permafrost has a very low hydraulic conductivity and commonly acts as a barrier to deeper groundwater recharge or as a confining layer to deeper aquifers (Ye et al. 2009). Because it is a barrier to recharge, permafrost increases the surface runoff and decreases subsurface flow. Permafrost extent over a region plays a key role in the distribution of surface–subsurface interaction (Carey and Woo 2001; Woo et al. 2008a). Permafrost and non-permafrost rivers have very different hydrologic regimes. Relative to non-permafrost basins, permafrost watersheds have higher peak flow and lower base flow (Woo 1986; Kane 1997). In the permafrost regions, watersheds with higher permafrost coverage have lower subsurface storage capacity and thus a lower winter base flow and a higher summer peak flow (Woo 1986; Kane 1997; Yang et al. 2003). It is a challenge to accurately determine changes in permafrost conditions. Our understanding of permafrost change and its effect on hydrological regime is incomplete. For instance, there are uncertainties regarding the impact of ground ice melt and its contribution to annual flow changes over large Siberian rivers (McClelland et al. 2004; Zhang et al. 2005a). Permafrost condition and streamflow characteristics vary within large watersheds in the northern regions. Examination and comparison of hydrological regimes between subbasins with various permafrost conditions can improve our understanding of impact of permafrost changes on cold region hydrology. Areas covered in this chapter are flow regimes in permafrost watersheds and the relationship with basin water storage (or terrestrial water storage) as well as permafrost coverage, base flow change and its indication for permafrost thawing, permafrost groundwater age, and basin water balance in the northern river basins.

16.2 Basin Permafrost Coverage and Flow Regime

Ye et al. (2009) examined the relationship between hydrological regime and permafrost coverage over nested subbasins within the Lena River in eastern Siberia. They analyzed monthly discharge data, with a focus on the ratio of the maximum to minimum discharge (Q_{\max}/Q_{\min}) and its relation with permafrost condition, because this ratio reflects the hydrological regime. The aim of their study is to quantify the impact of permafrost on streamflow regime and change, and to specifically define a relationship between basin permafrost extent and streamflow conditions over the Lena watershed. Ye et al. (2009) also examined relationship between basin air temperature and precipitation and their effects on permafrost extent and basin streamflow regimes. The results of their study enhance our knowledge of cold region hydrology and its change due to climate impact and human influence.

The Lena River originates from the Baikal Mountains in the southern part of central Siberian Plateau and flows northeast and north, entering into the Arctic Ocean via the Laptev Sea (Fig. 16.1). Its drainage area is about 2,430,000 km², mainly

covered by forest and underlain by permafrost. The Lena River contributes 524 km³ of freshwater per year, or about 15% of the total freshwater flow into the Arctic Ocean (Yang et al. 2002; Ye et al. 2003). Relative to other large rivers, the Lena basin has less human activities and much less economic development (Dynesius and Nilsson 1994). There is only one large reservoir in the Vilui subbasin. A large dam (storage capacity 35.9 km³) and a power plant were completed in 1967 near Chernyshevskyi (112.15 °E, 62.45 °N). This reservoir is used primarily for electric power generation: holding water in spring and summer seasons to reduce snowmelt and rainfall floods and releasing water to meet the higher demand for power in winter (Ye et al. 2003). Various type of permafrost exists in the Lena basin, including sporadic, or isolated permafrost in the source regions, and discontinuous and continuous permafrost in downstream regions (Fig. 16.1) (Brown et al. 1997). Approximately 78–93% of the Lena basin is underlain by permafrost (Zhang et al. 1999; McClelland et al. 2004). An overview of the geographical scope of the Lena River basin and the seasonal changes and long-term trends of the Lena River discharge are also described in detail by Hiyama et al. (2019).

Monthly discharge regimes vary over the Lena basin, although it is generally high in summer and low in winter. The ratio of Q_{\max}/Q_{\min} is a direct measure of hydrologic regime. They increase with drainage area from the headwaters to downstream within the Lena basin. This pattern is different from the non-permafrost watersheds. The ratios also change with basin permafrost coverage, suggesting that permafrost affects regional hydrological features. Regression analysis between the Q_{\max}/Q_{\min} ratio and basin Coverage of Permafrost (CP) reveals a significant relationship at 99% confidence. This relationship quantifies the effect of permafrost distribution on discharge process. It shows that permafrost conditions do not significantly affect streamflow regime over the low permafrost (less than 40%) regions, and it strongly

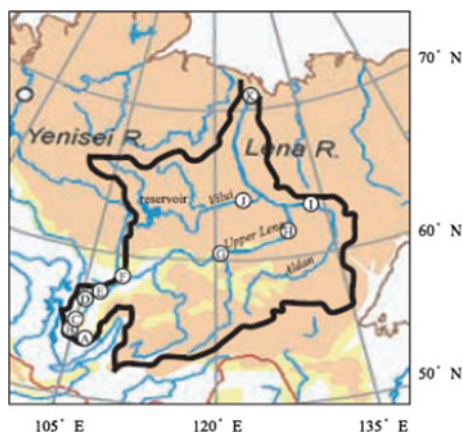


Fig. 16.1 The Lena River watershed permafrost distribution and hydrological stations (A through K) used in Ye et al. (2009)

affects discharge regime for regions with high (greater than 60%) permafrost coverage (Fig. 16.2). It is important to define this relation of permafrost impact to regional hydrology, as this knowledge is useful for further investigation into permafrost effect on basin hydrology over the large northern regions. On the other hand, temperature (T) and precipitation (P) have similar patterns among the subbasins. Basin precipitation has little association with permafrost conditions. There is a reasonable relationship between the freezing index and permafrost extent over the basin, indicating that cold climate leads to high coverage of permafrost. This relationship is useful to examine basin thermal condition with permafrost distribution. The combination of the relations between T–CP and CP– Q_{\max}/Q_{\min} links temperature, permafrost, and flow regime over the Lena basin. The variations in the Q_{\max}/Q_{\min} ratios reflect hydrologic regime changes. Trend analyses of the ratios show different results over the Lena basin. The ratios significantly decrease in the Vilui valley and at the Lena basin outlet due to dam regulation. In the unregulated upper Lena regions, the ratios do not show any trend during 1936–1998. This may imply that the permafrost changes in the region were not sufficient to affect hydrological processes. In the Aldan tributary without regulation, the ratios significantly decrease due to increase in base flow during 1942–1998. This change is consistent with significant group temperature warming (Fedorov et al. 2014a) and permafrost degradation over the eastern Siberia (Fedorov et al. 2014b). This consistency to some extent verifies permafrost influence on basin streamflow process and change (Ye et al. 2009).

Yang et al. (2014) recently used various statistical approaches for data analyses to calculate the mean, and standard deviation of the long-term daily discharge records of the Mackenzie River. Relative to other seasons, the minimum daily flows (usually in the winter season) range from 1680 to 4090 $\text{m}^3 \text{s}^{-1}$ over the study period. The interannual variation is much smaller than that for the maximum daily flows. There is a weak tendency in the minimum daily flow increase during 1973 to

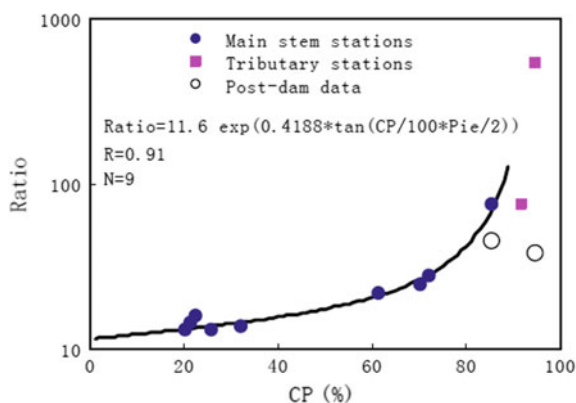


Fig. 16.2 The ratios of Q_{\max}/Q_{\min} versus Coverage of Permafrost (CP) at the nine stations from the upper Lena to basin outlet. Two additional stations (the Aldan and Vilui subbasins) are also shown Ye et al. (2009)

2011. This change in flow amount is very small and statistically insignificant; it is thus almost undetectable in terms of its contribution to the total flow, i.e., stable ratios of minimum daily flow/mean daily flow over the study period. It is important to mention that base flow increase has been reported for the northern regions and the Northwest Territories (NWT), Canada, due to recent climate warming (Woo et al. 2008a, b; Woo and Thorne 2014; St. Jacques and Sauchyn 2009). Yang et al. (2014) discovered Mackenzie basin monthly discharge increases during September to April. Basin water storage changes affect low-flow conditions and their changes. The Mackenzie basin has large seasonal storages due to many large lakes in the basin. The increases in the monthly and daily low flows may reflect changes in the basin storages, including lakes, groundwater, and permafrost and ground ice. Walvoord and Striegl (2007) also reported an increase in groundwater to stream discharge from permafrost thawing in the Yukon River basin. Suzuki et al. (2018) compared river discharge (R) and the GRACE (Gravity Recovery and Climate Experiment)-derived Terrestrial Water Storage (TWS) in the three largest pan-Arctic river basins from April 2002 to December 2016. They revealed strong positive correlations between R and TWS for the Lena River and Mackenzie River basins, but no significant correlation for the Yukon River basin, where TWS was sensitive to the changes in evapotranspiration. They also detected that the TWS among the three river basins was further controlled by the presence of continuous permafrost. Namely, the autumnal TWS in the Lena River basin, which exhibits continuous permafrost, persisted R through the spring of the following year. However, no such effect was observed in the Mackenzie River catchment, which is partially covered in both continuous permafrost and discontinuous permafrost. These results suggest regional to large scale change in permafrost characteristics. There is therefore a need to continue to explore the relationship among climate, river flow, and permafrost conditions over the high-latitude regions.

16.3 Detection of Permafrost Thawing Trends from Long-Term Streamflow Measurements

The low flows in a basin generally are fed by groundwater seepage from the upstream riparian aquifers in the basin. In the cold regions, permafrost thawing and permafrost growth directly affect the groundwater storage amount and mobility, which in turn control the seepage from these aquifers. Thus, long-term streamflow records are a useful but hitherto unexplored source of information on the past history of permafrost changes during the same period. On the basis of this principle and hydraulic groundwater theory, Brutsaert and Hiyama (2012) proposed three methods to relate low flows during the open water season with the rate of change of the active groundwater layer thickness resulting from permafrost thawing at the scale of the upstream river basin. They applied their methods to detect basin-scale permafrost thawing trends from long-term summertime base flow measurements.

16.3.1 Base Flow and Basin-Scale Groundwater Storage

The proposed approach is an extension of a method developed earlier (Brutsaert 2008, 2012; Brutsaert and Sugita 2008) to determine groundwater storage changes, as manifested in available streamflow records. Streamflow occurring in the absence of precipitation or upstream water inputs is referred to variously as base flow, fair weather flow, drought flow, or low flow; it can normally be assumed to consist of the cumulative outflow from all upstream phreatic aquifers along the banks. The lowest of the low-flow recessions, as small perturbations of the no-flow steady state, can simply be expressed as an exponential decay phenomenon and can be formulated as

$$y = y_0 \exp(-t/K) \quad (16.1)$$

where y is the rate of flow in the stream per unit of catchment area, y_0 is the value of y at the arbitrarily selected time origin $t = 0$, and K is the characteristic time scale of the catchment drainage process, also commonly referred to as the storage coefficient. Note that $0.693 K$ can be considered the storage half-life of the upstream riparian aquifers. The water stored in a river basin can be expressed as linear relationship between groundwater storage $S = S(t)$ and base flow $y = y(t)$, as

$$S = Ky. \quad (16.2)$$

The storage S can be visualized as the average thickness of a layer of water above the zero flow level. However, the thickness of this layer occupied inside the soil profile is larger than S . In the simplest approach, capillary effects above the water table are parameterized by means of the drainable porosity n_e , also known as the specific yield. This means that the water table is assumed to be a free surface and the fraction of the soil volume occupied by free and movable water is assumed to be n_e . The average thickness of the water layer stored in the soil profile, that is the active groundwater layer above permafrost, is given by

$$\eta_0 = S/n_e. \quad (16.3)$$

Before using (16.2) and (16.3) to relate changes in average layer thickness η_0 with observed changes in base flow y , it is necessary to determine how changes in η_0 might affect the drainage time scale parameter K .

16.3.2 Active Groundwater Layer Thickness Trend

If it can be assumed that n_e in the aquifer remains unaffected by changes in the underlying permafrost layer, the determination of the growth of the groundwater active layer can be expressed as follows:

$$\frac{d\eta_0}{dt} = \frac{K}{2n_e} \frac{dy}{dt}. \quad (16.4)$$

If a good estimate of the drainable porosity n_e and of the drainage time scale K is unavailable or impossible, but an estimate of a reference or typical value of the layer thickness, say η_{0r} , can be obtained, an alternative expression can be derived.

$$\frac{d\eta_0}{dt} = \frac{\eta_{0r}}{2y_r} \frac{dy}{dt} \quad (16.5)$$

where y_r is a typical or reference base flow around the time when η_{0r} was observed. The choice between (16.4) and (16.5) has to depend on the availability of estimates of K , n_e , and η_{0r} .

Whenever an estimate of a reference or typical value of the active groundwater layer thickness η_{0r} is available, and because the drainage time scale K is inversely proportional to η_0 , the growth rate of the active groundwater layer can also be written as

$$\frac{d\eta_0}{dt} = \eta_{0r} K_r \frac{dK^{-1}}{dt} \quad (16.6)$$

where K_r is the value of the drainage time scale at the time of η_{0r} .

16.3.3 Application to the Lena River Basin

The three methods were tested with data from four gaging stations within the Lena River basin in eastern Siberia. The data were obtained from two sources. The daily discharge data for the period 1950–2003 were obtained through courtesy of the Japan Agency for Marine-Earth Science and Technology (JAMSTEC), whereas for the more recent period 2000–2009, they were obtained from R-ArcticNet (A Database of Pan-Arctic River Discharge developed by the Water Systems Analysis Group at the University of New Hampshire). The three methods were applied with the streamflow data of May through September for each decade from 1950 through August 2009 for the four subbasins (upper Lena, Olyokma, and two Aldan subbasins) in the Lena River watershed.

Because, sometimes, daily flow data are subject to measurement uncertainties, it was decided to use a running average of five days as a more reliable measure of y . These annual lowest 5-day flows were calculated for each of the four basins, and the resulting values were then used to calculate the changes in base flow dy/dt , which are needed for (16.4) and (16.5). Beside dy/dt , a knowledge is required for the characteristic drainage time scale K and of its long-term changes dK^{-1}/dt or dK/dt . The procedure consists of determining the lower envelope of a logarithmic plot of appropriate $-dy/dt$ data versus the corresponding y data; it also makes use of the observations for $y = -K(dy/dt)$.

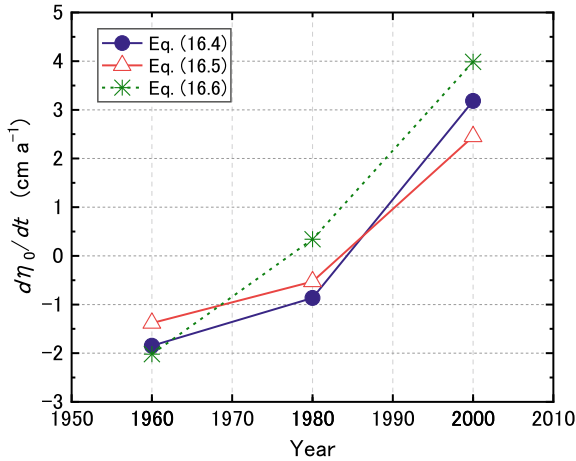


Fig. 16.3 Trends of the active groundwater layer thickness $d\eta_0/dt$ in cm a^{-1} over the periods 1950–1969, 1970–1989, and 1990–2008 in the Aldan subbasin (Aldan-2 (A2) of Brutsaert and Hiyama 2012). Needed values for the use of Eq. (16.4) were $n_e = 0.01$ and $K = 46.33$ as the station average. Those for Eq. (16.5) were $\eta_{0r} = 1.88$ m and y_r as average of the annual lowest 5-day flows during each period. Those for Eq. (16.6) were $\eta_{0r} = 1.88$ m and $K_r = 46.33$ as the station average

In the three expressions to calculate $d\eta_0/dt$, namely (16.4), (16.5), and (16.6), it would seem that (16.5) is the most parsimonious one because only required knowledge or input is the average active groundwater layer depth η_{0r} at the end of summer. Based on the ground temperature measurements during 1956–1990 at 17 stations over the Lena River basin, Zhang et al. (2005b) found the average active layer thickness to be 1.88 m, ranging from 1.2 m to 2.3 m. With the assumptions that $\eta_{0r} = 1.88$ m, and that y is the average of the annual lowest 5-day flows during each period, it is possible to implement (16.5). The results of this application for the Aldan subbasin are presented in Fig. 16.3.

Equation (16.4), similar to (16.5), can also be implemented by means of the values of dy/dt with the average of the annual lowest 5-day flows. In addition, however, values of K and of n_e are required. No information is available for the drainable porosity n_e in the Lena region. On the basis of several findings (e.g., Brutsaert 2008) and in the absence of more specific information of n_e , (16.4) was implemented with the values of $n_e = 0.01$, dy/dt using the average of the annual lowest 5-day flows during each period, and K averaged over each of the three periods. The results are presented and compared in Fig. 16.3 with that from (16.5) so as to check how realistic the adopted value n_e is. The results from (16.4) can match that from (16.5) by very slightly adjusting n_e from 0.01 to 0.0121. This is a very small change that supports the use of the present rough estimate of $n_e = 0.01$.

For the application of (16.6), the $d(K^{-1})/dt$ values were calculated as finite differences between the values of K^{-1} at different times. The required values of K_r

were taken as the average values for each basin, and the corresponding value of the active groundwater layer thickness was given as $\eta_{0r} = 1.88$ m. Relative to the results obtained with (16.4) and (16.5), (16.6) produced higher $d\eta_0/dt$ in general (Fig. 16.3). It is difficult to judge or assess which method had the best performance. Due to subjective determination of K and a fortiori its derivatives $d(K^{-1})/dt$ for (16.6), and the relatively straightforward way of determining dy/dt , it seems that (16.4) and (16.5) should be given preference over (16.6). Because (16.5) makes use of measured data, it should probably be given preference over (16.4).

Brutsaert and Hiyama (2012) also compare the values of $d\eta_0/dt$ with the recent field observations, which indicate acceleration in permafrost deterioration. At the study site of Spasskaya Pad, some 20 km north of Yakutsk, the annual maximum depth of the 0 °C isotherm was recorded (Ohta et al. 2008) to increase from around 1.27 m in 1998 to 2.0 m in 2006, i.e., an average rate of about 9.1 cm a⁻¹. On the other hand, in a parallel study at the same location (Iijima et al. 2010), manual observations of the annual maximal thaw depth by a frost tube indicated that between 2000 and 2007 it increased from around 1.37–1.67 m, i.e., at a rate of about 3.3 cm a⁻¹. Since manual measurements are considered more reliable, the latter rate of 3.3 cm a⁻¹ is probably more representative. This change is in agreement with the results shown in Fig. 16.3.

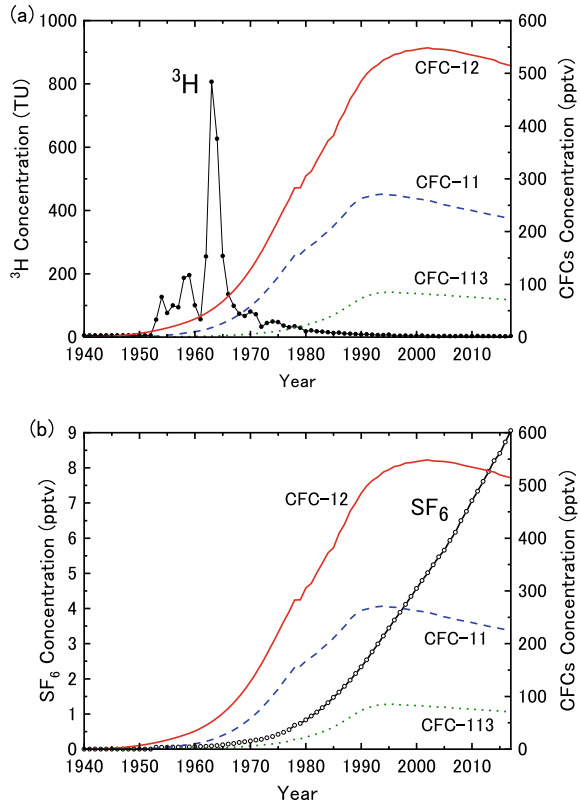
16.4 Determination of Permafrost Groundwater Age

Understanding groundwater characteristics such as pathways and residence time (groundwater age) is the most important research issue in groundwater hydrology as well as for sustainable use of groundwater resources. This issue is also applicable for permafrost region. Transient tracers such as tritium (³H), chlorofluorocarbons (CFCs), and sulfur hexafluoride (SF₆) have been widely used to determine groundwater age in humid and semi-arid regions (Busenberg and Plummer 1992, 2000; IAEA 2006). However, only a few studies have used such tracers in permafrost regions. Hiyama et al. (2013) is one of such researches, and they have applied transient tracers to detect groundwater age to several spring discharges in eastern Siberia. In this section, the paper of Hiyama et al. (2013) is introduced and revisited very briefly.

16.4.1 Groundwater Age Estimation Using ³H

Tritium (³H) concentration has previously been used for estimating groundwater age, especially in the Northern Hemisphere. The ³H concentration of precipitation peaked around 1963 due to nuclear testing (Fig. 16.4a). The half-life of ³H is relatively short (12.3 years), and therefore when nuclear testing was terminated the concentration declined. However, if the mean residence time of permafrost

Fig. 16.4 Time series of the tritium (^3H) concentration (TU) in precipitation **a** obtained at Ottawa, Canada, observed by IAEA (International Atomic Energy Agency) (<https://nucleus.iaea.org/wiser/index.aspx>). Also shown are the atmospheric CFCs concentrations (**a**, **b**) and the atmospheric SF_6 concentration **b** observed by the Reston Groundwater Dating Laboratory of the USGS (US Geological Survey) (https://water.usgs.gov/lab/software/air_curve/index.html)



groundwater is around 50–60 years, ^3H may be a useful tracer to detect the extent to which the groundwater originated from precipitation after nuclear testing.

16.4.2 Groundwater Age Estimation Using CFCs and SF_6

CFCs and SF_6 , as transient tracers, can also be used to estimate groundwater age (IAEA 2006). CFCs are anthropogenic gases that were emitted into the atmosphere from the 1930s to 1987. Atmospheric concentrations of the major CFCs such as dichlorodifluoromethane (CCl_2F_2 ; CFC-12), trichlorofluoromethane (CCl_3F ; CFC-11), and trichlorotrifluoroethane ($\text{C}_2\text{Cl}_3\text{F}_3$; CFC-113) peaked in the 1990s (Fig. 16.4). SF_6 emissions began in the 1960s as it was used as an isolation gas, and its use continued through recent decades (Fig. 16.4b). Thus, CFCs can be used to estimate groundwater age recharged around 30–80 years ago. SF_6 can also be used for groundwater recharged around 1–50 years ago.

Three assumptions are needed when we apply CFCs and SF_6 concentrations to estimate groundwater age. First, the sampled water truly contained CFCs and SF_6 concentrations of the target groundwater in the aquifer. Second, the recharged water

reached full equilibrium in solubility with the soil gas (or the atmosphere), and the dissolved CFCs and SF₆ were preserved within the aquifer. Third, ground ice melting in permafrost contributes to decline the concentrations of CFCs and SF₆ of the aquifer. This is because the above-mentioned assumptions could be also applied to the permafrost groundwater layer.

Henry's solubility law (e.g., Warner and Weiss 1985) is used to estimate recharge year (when the sampled water entered the aquifer) from the dissolved CFCs and SF₆ in the sampled water. For accurate chemical analyses of dissolved CFCs and SF₆ concentrations in the sampled water, they should be kept isolated from the ambient atmosphere. Using analyzed concentrations of CFC-12, CFC-11, CFC-113, and SF₆ from the sampled spring water together with temperature and elevation data, it is possible to estimate the recharge year using Henry's solubility law.

16.4.3 Application to the Lena River Basin

Spring water samples were taken in July 2009, July 2010, August 2011, and August 2012, at four sites: Buluus (61°20'N, 129°04'E), Ulakhan-Taryn (61°34'N, 129°33'E), Eruu (61°42'N, 129°45'E), and Michita (61°23'N, 129°11'E) discharges. These four spring discharges are hydrological monitoring sites of the Melnikov Permafrost Institute of the Siberian Branch of the Russian Academy of Sciences.

³H counting was conducted using a low-background liquid scintillation counter, Aloka model LB5, following electrolytic enrichment of ³H by a factor of about 25 using Fe–Ni electrodes. Total analytical precision was better than ±0.23 tritium unit (TU) (±1σ). The ³H measurements were conducted at the Geo-Science Laboratory Co. Ltd, Nagoya, Japan. CFC content in the samples was measured using a purge and trap Gas Chromatography procedure with an Electron Capture Detector (GC-ECD) at the Geo-Science Laboratory Co. Ltd, Nagoya, Japan. The procedure involved stripping 40 mL sample water of CFCs using ultra-pure nitrogen gas. The extracted CFCs were purified and concentrated using a cold trap, and finally injected into the GC-ECD. The precision and detection limit of the analysis were less than 2% and 1 pg L⁻¹, respectively. SF₆ content was measured using the same procedure and GC-ECD at the Geo-Science Laboratory. The only difference was that, after stripping 400 mL sample water, the extracted SF₆ was purified and concentrated using two cold traps. The precision and detection limit of the analysis were less than 3% and 0.05 fmol L⁻¹, respectively.

³H was detected in all of the spring water. The concentration at Buluus and Eruu springs was higher (9–11 TU) than that at Ulakhan-Taryn spring (1–7 TU). ³H concentration at Michita is the highest (15–17 TU). The clear difference among the springs indicates that the origin and mixing conditions of groundwater are different. There was no significant difference in ³H concentration at Buluus and Michita in the sampling years (2009–2012). Buluus and Eruu springs may have mostly originated from precipitation that recharged the aquifer after the nuclear testing period of the 1960s. If we assume that groundwater pathways in talik zone have a piston-like

flow, the ages of the Buluus and Eruu spring water can be estimated to be about 1–40 years old. However, because the ^3H concentrations of both samples were not significantly higher than those of precipitation at Yakutsk, the water of Buluus and Eruu may contain older precipitation recharged before nuclear testing. Ulakhan-Taryn was the lowest value compared to Buluus, Eruu, and Michita. As ^3H concentration of the Ulakhan-Taryn samples was clearly lower than that of precipitation that fell after the 1960s, it may have been recharged mostly by precipitation before the 1960s. If we set the ^3H concentration of precipitation after the 1960s at 16 TU (the current value at Yakutsk), and that before the 1960s as 0.5 TU (the concentration of the 1950s), the percentages of precipitation recharge before the 1960s to the Ulakhan-Taryn spring water are 90–95%. If we assume that groundwater pathways have a piston-like flow, the recharge year of the spring water can be estimated as 1955, i.e., the age of the groundwater is 55 years. The Michita water seems to have been recharged by very recent precipitation.

CFC-12 was detected at all sites. However, CFC-113 was not detected at Ulakhan-Taryn and Eruu springs. This may have been caused by deoxidizing conditions at Ulakhan-Taryn and Eruu. Under deoxidizing conditions, methane-producing bacteria appear to remove dissolved CFC-11 and CFC-113 (Happell et al. 2003). In contrast, three kinds of CFCs (CFC-12, CFC-11, and CFC-113) were detected at Buluus and Michita. Thus, microbial degradation likely plays a smaller role at both springs. This means the groundwater age could be estimated more reliably at Buluus and Michita.

The concentration of SF_6 ranged from 1.71 to 5.67 fmol kg^{-1} , and was higher at Michita and Buluus than at Ulakhan-Taryn and Eruu. This result is similar to that of the CFCs analysis, suggesting that the groundwater age of the Buluus and Michita springs is younger than that of the Ulakhan-Taryn and Eruu springs.

The equivalent air concentration of the three CFCs and the SF_6 was calculated using Henry's solubility law (Warner and Weiss 1985) in which the solubility potentials of CFCs and SF_6 depend on the air temperature and air pressure when the precipitation fell. Because the water should be recharged above 0 °C, representative value of the temperature was assumed to be 0 °C. Mean surface elevation of the recharge area was assumed to be 150 m based on the topography map. The estimated recharge years using the CFCs method were from around 20 to 40 years. However, those using the SF_6 were quite younger. The SF_6 concentration might have increased mainly due to the mixture of SF_6 from basement rock with permafrost groundwater, which would cause a younger estimate of SF_6 -based groundwater age. Another possible reason is due to excess air, redundant SF_6 may have dissolved into the recharging water, leading to overestimated SF_6 levels and thus younger groundwater ages. Although these differences were observed among three tracers (^3H , CFCs, and SF_6), the relative order of apparent ages was consistent with each other.

Finally, we should note that the ^3H -based ages and the CFC-12-based ages at Ulakhan-Taryn springs varied widely. This is because we applied a simple piston-like flow scheme to the groundwater system, although the samples were a mixture of supra-permafrost and intra-permafrost groundwater. It was not possible

to separate these two types of permafrost groundwater. Nevertheless, it should be noted that large variability in the concentrations of ^3H and CFC-12 at Ulakhan-Taryn indicates unstableness of groundwater pathways and its recharge sources. Because soil pore ice was found below the active layer, i.e., between the active layer and the upper boundary of the intra-permafrost groundwater aquifer in this area, melted pore ice could contribute to the source of the spring water at Ulakhan-Taryn.

16.5 Water Balance of Northern Permafrost Basins

Water balance analysis is a widely used tool for evaluating the availability and sustainability of water resources and supply (Healy et al. 2007). Carried out for a long enough period, the results of water balance calculation can be useful for quantifying natural variability, climate change, and man-induced alterations. Ice, snow, frozen soils, and large annual fluctuations in the surface energy balance are characteristics of the high-latitude watersheds. The hydrologic responses of northern watersheds differ from those in the temperate regions, and such differences should be reflected in the water balance results. The difficulty in quantifying these differences is mainly due to few water balance studies at high latitudes, often marginally supported and only for short study periods.

Most of the early studies concentrated on watersheds in areas of discontinuous permafrost because of logistical reasons. To carry out water balance computations, it is necessary to measure liquid and solid precipitation and runoff and to make estimates of evapotranspiration losses and any significant changes in storage. All these elements of the hydrologic cycle in the high latitudes are quite variable and often are not accurately measured.

Water balance data from 39 high-latitude watersheds (Fig. 16.5) over the past 50 or more years were presented and discussed in a synthesis workshop in 2004 at Victoria, BC, Canada (Kane and Yang 2004a). Less than one-third of these watersheds continue to be active. These studies ranged from one year in length to over 50 years (Seuna and Linjama 2004). Characteristically, these experimental watersheds are small, from less than 1 km² to 432 km².

The three main features of the water balance equation are the input (P), changes in various storage terms (DS), and outputs (ET and R):

$$P = ET + R \pm DS_{\text{sur}} \pm DS_{\text{sub}} \pm DS_{\text{vad}} \pm DS_{\text{glac}} \pm DS_{\text{snow/ice}} + \eta \quad (16.7)$$

where P is precipitation, including both snowfall and rainfall (and possibly also condensation), ET is evapotranspiration (possibly also sublimation), R is discharge throughout the entire period of flow, DS_{sur} is change in surface storage (lakes, wetlands, reservoirs, channels, etc.), DS_{sub} is change in subsurface storage of groundwater, DS_{vad} is change in storage of unsaturated (vadose or active layer) zone, DS_{glac} is change in glacier storage, $DS_{\text{snow/ice}}$ is change in storage of

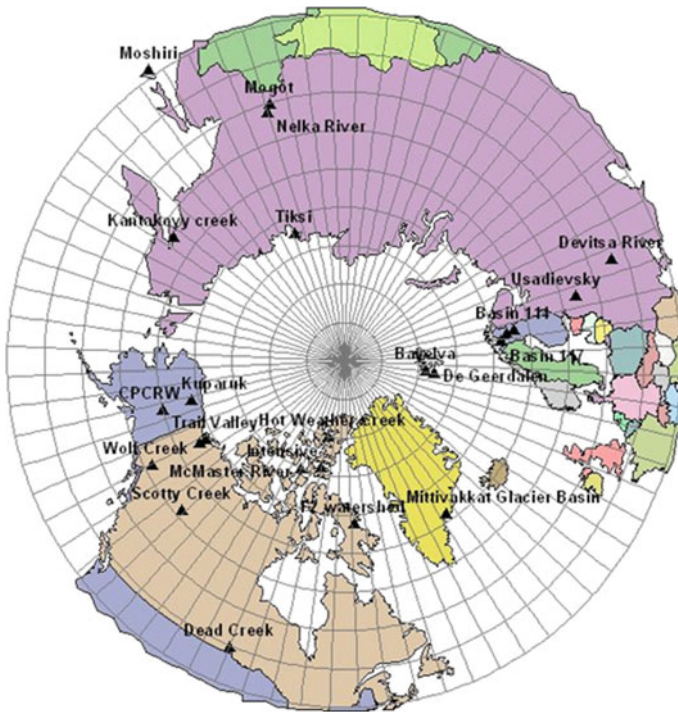


Fig. 16.5 Names and locations of 39 long-term water balance study watersheds in the northern regions. (map modified from Kane and Yang (2004a))

snowpack (including late-lying snowdrifts, aufeis, river and lake ice, etc.), and η is error term on closure. Typically, precipitation and discharge are measured by national hydrologic networks. However, in extreme environments, the error associated with these measurements can be considerable (e.g., the difficulties in measuring solid precipitation and streamflow in ice-choked drainages).

Except for experimental watersheds, all of the terms in this equation are seldom measured. ET is difficult to spatially and temporally quantify over a watershed through field measurements. Even for experimental watersheds, changes in the various storage terms are challenging to quantify; more often they are assumed to remain unchanged over a time period of a year (although this is seldom true). Theoretically, η should be zero, a condition seldom satisfied. For those watersheds where the error term was determined (Kane and Yang 2004b), the error ranged from zero to 27% of annual precipitation, with an average of about 8%. Assuming that the quality of measurements are the same in all research watersheds, this implies that overall researchers are doing a fair job of measuring the variables in the water balance equation, but it does not indicate the measurement error associated with each individual term. Although there are many limitations to these water balance

studies, they represent the best hydrologic data that we presently have for northern basins. Therefore, results from the IAHS Redbook papers (Kane and Yang 2004a) on the individual watersheds and the synthesis papers should be useful for verifying modeling results from meso- and global-scale climate models. For example, modeled fluxes should correspond reasonably close to measured fluxes from these individual watershed studies for a given region. This could be extended to estimate the latitudinal decrease in the precipitation rate based on the data obtained collectively from all watersheds, particularly for basins dominated by a continental climate. For the watersheds with a continental climate, there is a decreasing rate in annual precipitation of 16 mm/8 latitude (Kane and Yang 2004b). A similar decreasing trend is seen for more coastal watersheds, but their annual precipitation rates are higher.

On average, for all of the watersheds, about half of the annual precipitation leaves the basin as ET. Although there are some outliers, the decreasing trend in ET with latitude is numerically similar to that of precipitation. This trend is primarily driven by declining solar radiation at the surface (also reflected in the lower mean annual temperature) and in some cases by moisture limitation (e.g., in the Canadian High Arctic). Methods of measuring/estimating ET are quite variable, ranging from mass and energy balances to simple empirical methods (Kane et al. 1990; Mendez et al. 1998; Shutov et al. 2006). Owing to the wide variety of methods for obtaining ET and because of their disjointed measurement periods, it is difficult to elucidate its spatial and temporal variability. An area of pressing interest is to improve estimates of ET, particularly for larger areas.

In this context, newly produced reanalysis data are useful for detecting spatial and temporal variability of ET. Using multi-model ensemble of reanalyzed data from GLDAS1 and GLDAS2, Suzuki et al. (2018) revealed negative trend in the TWS throughout the Arctic Circumpolar Tundra Region (ACTR) and this was primarily driven by ET. Meanwhile, it was indicated that precipitation has a minor impact on the TWS.

The average percentage of annual precipitation is falling as snow increases with latitude from a low of about 10% to a high of about 80%. Again, there is considerable variation, particularly for coastal watersheds. It is apparent that the snow water equivalent of the annual snowpack for most watersheds, with the exceptions of some near the coast, is generally below 200 mm a⁻¹ (Kane and Yang 2004b). Ideally, we should have several research watersheds in the circumpolar Arctic that have been in operation for several decades and are much larger than those described in Kane and Yang (2004b). This would allow us to capture the natural variability (both drought and flood conditions) and to look for trends in the fluxes and storage terms due to climate change. The present experimental watersheds are so small that when considered in the global context, they are no more than single points. It is also clear that we need to do a better job of actually measuring the various fluxes and storage reservoirs.

16.6 Summary and Future Directions

This chapter reviewed and discussed selected topics in permafrost hydrology of the northern regions, particularly the association of Lena river flow regime with permafrost coverage within nested watershed, base flow change and its implication to active layer variation, groundwater age, and water balance analyses in long-term experimental basins across the northern regions. Below are a brief summary of main discussions and some consideration for future research needs.

There are similarity and variation in streamflow regimes over the Lena watershed. The ratios of monthly maximum/minimum flows directly reflect discharge regimes. The ratios increase with drainage area from the headwaters to downstream within the Lena basin (Ye et al. 2009). It is important to point out that pattern is different from the non-permafrost watersheds, and it clearly reflects permafrost effect on regional hydrological regime. There is a significant positive relationship between the ratio and basin permafrost coverage. This relationship indicates that permafrost condition does not significantly affect streamflow regime over the low permafrost (less than 40%) regions, and it strongly influences discharge regime for regions with high permafrost (greater than 60%). This is consistent with the result of Suzuki et al. (2018). There is a need to refine this relationship for various climate and permafrost conditions in the high-latitude regions.

Three distinct procedures were developed by Brutsaert and Hiyama (2012) to relate summer low flows with the changes in the active groundwater layer thickness due to permafrost thawing in the upstream Lena river basin. Their methodology has distinct benefits: first, it provides a tool to upscale local measurements over the entire upstream river basin, possibly after calibration with such local measurements; second, it can produce historical information of permafrost conditions that predate most field observations since streamflow measurements started well before permafrost observations. Application of the methods in the Aldan subbasin of the Lena River show that, in the period 1950–1970, active layer thickness decreased without much thawing. However, from the 1990s onward, active layer growth rates were as high as 2 cm a^{-1} or more. These results agree with previously published active layer changes at many sites in eastern Siberia. However, accurate determination of the drainage time scale K and its temporal rate of change dK/dt or $d(K^{-1})/dt$ still poses challenges. Additionally, it should be noted that the four catchments in Brutsaert and Hiyama (2012) are too large to draw strong conclusions. It is thus recommended that the methods should be applied to smaller basins for validations. The loss of upper permafrost layer might be more serious than what Brutsaert and Hiyama (2012) estimated since they assumed that the entire catchment contributed to the base flow, whereas in reality the contributing area is limited to the immediate vicinity of the river channel.

The apparent ages of the mixture of supra-permafrost and intra-permafrost groundwater in the middle of the Lena River basin were estimated to be 1–55 years old. The spring water at Ulakhan-Taryn appeared to contain more than 90% recharge by precipitation before the 1960s nuclear testing era. Because soil pore ice

was found below the active layer around the site, melted ice could contribute to the source of the spring water. If recent global warming causes permafrost degradation in the region, it could drive changes of groundwater recharge source as well as the rates, especially increasing the contribution from thawing permafrost. It is important to note the insufficient data to characterize permafrost characteristics and groundwater conditions in cryohydrogeological point of view. Groundwater dating is one of the greatest challenges in permafrost regions.

There has been increased attention at the permafrost regions hydrology due to resource development and climate change over the northern regions. This chapter reflects some recent advances in permafrost hydrology. Much of the recent research has focused on surface water, less on springs and groundwater contribution to streamflow. It is very clear that data limitation and poor quality remain serious concerns for permafrost hydrology research and applications. A compendium of water balance research from 39 high-latitude catchments reveals the strengths and limitations of the available results, as most of the basins are restricted to only a few years of study at the small watershed scale. The response of discharge to climate receives increasing attention, including extreme hydrologic events to the changing flow regimes in large northern watersheds. The effect of climate change and the role of permafrost on the changing discharge of large pan-Arctic rivers are major interest for further investigation. Extended field and modeling research on physical processes will improve knowledge of permafrost hydrology and enhance its relevance to societal needs.

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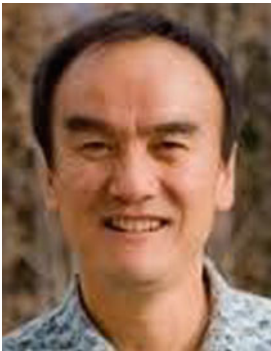
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