Permafrost Features and Talik Geometry in Hydrologic System

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Abstract

Permafrost is widely distributed in the high latitudes. This chapter discusses frozen (permafrost) and unfrozen states of the hydrological geometry in the northern regions. The hydrological activities are very active and dynamic not only in discontinuous permafrost zone but also in cold continuous permafrost areas. Water carries significant amount of heat in aquifer and talik system. Water locates in the depth below the maximum ice formation can develop an unfrozen layer underneath the water body (i.e., talik and thaw bulb). Taliks could be open to connect to the sub-permafrost layer, while the hydrologic gradient makes flow in upward or downward directions. The heat balance of the super-, inter-, or sub-permafrost generates unique unfrozen geometry in the permafrost. This chapter also reviews various cellars developed and used in the arctic regions by indigenous people.

14.1 Introduction

Presence or absence of the permafrost have been discussed for years in the Arctic in regard to understanding the hydrological cycle in the cold environments. In general, permafrost acts as an impermeable substrate, that is why there are so many lakes, ponds, and wetlands in areas of permafrost regions despite the low amounts of

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D. Yang and D. L. Kane (eds.), *Arctic Hydrology, Permafrost and Ecosystems*,
https://doi.org/10.1007/978-3-030-50930-9_14

annual precipitation. However, the property of hydraulic conductivity is not universally low; it can vary considerably depending upon soil/rock type, permafrost temperature, salinity, etc. Thin films of unfrozen water exist on the surface of the porous media and the greater the surface area of the substrate, the thicker the unfrozen film (greatest for warm clays). Generally, permafrost is considered as impermeable for hydrologic applications with sub-permafrost groundwater below the permafrost and suprapermafrost groundwater above.

In this chapter, we define terms of some permafrost terms and discuss the interaction between frozen soil and water. Permafrost is defined as any ground material (soil, rock, ice, water) staying below 0 °C for two or more consecutive years by today's scientific community and the western world, some of the former Soviet countries and Eastern countries define it as several years instead of two. However, these definitions prove to be problematic as not all water in permafrost freezes at 0 °C.

The current distribution of the permafrost is not only related to today's climate but also strongly linked to the late Pleistocene climate (e.g., glacier distributions) and somewhat by the Little Ice Age epic. The major permafrost forming period in the Arctic was slightly prior to the Last Glacial Maximum (LGM). That is why a majority of the ground ice (syngenetic ice wedges), also called "Yedoma" deposits, are present over most of the unglaciated regions. Also, permafrost is much more extensive in unglaciated areas, reaching depths exceeding 1 km in Eastern Siberia. Late Pleistocene glaciations covered considerable terrain in the North America, Scandinavia, and Greenland; these ice sheets blocked the severe climate from producing deeper permafrost (in some cases no permafrost). After retreat of the ice sheets, glacial isostatic rebounding took place in these areas and in one case producing newly developed coastal areas. This resulted in the development of young permafrost with complicated hydrogeological dynamics, including enrichments of the brines and the formation of unfrozen pockets (cryopegs).

Over the last fifty years, soil moisture measurement methods have been improved by the introduction of electromagnetic technology. Since the semiconductor industry has grown dramatically over the recent decades, many of the microchips (processing unit) and oscillators required for electromagnetic methods are now available at very low cost and small size. Cheaper, smaller, and more portable devices for measurement read-out (or datalogging) have also appeared on the commercial market. This has dramatically changed our understanding of active layer dynamics in Arctic regions. Smith and Tice (1988) compared nuclear magnetic resonance (NMR) and Time Domain Reflectometer (TDR) measurements of the unfrozen water content of frozen soils. The NMR technique is one of the most reliable methods for determining the unfrozen water content of soils (Tice et al. 1982), as it depends on the spatial density of hydrogen atoms in the sample. They concluded that very accurate readings were also possible using TDR for unfrozen water content determinations. Unfortunately, NMR's use is restricted to the laboratory. The water content of frozen materials strongly controls the freezing and thawing process (through latent heat exchange) of the active layer in arctic regions. The moisture content of the soil also affects the thermal properties (thermal offset) of permafrost that is directly related to the thermal conductivity (Yoshikawa et al. 2002). Thus, periglacial processes are strongly influenced by the water content of the materials.

Talik is originally a Russian term about year around unfrozen soil structure of supra-, inter- or sub-permafrost layer. Taliks occur in many situations involving brackish and fresh water and are very common to find in the Arctic. Many arctic communities use fresh water taliks as drinking water sources. Cryopeg is another Russian term for the layer of unfrozen ground where phase change is prevented by freezing point depression due to the dissolved solids content of the pore water. The freezing point of the cryopeg is interestingly close to the zero annual amplitude permafrost temperature (Yoshikawa et al. 2004). Cryopegs are not as domestically useful as freshwater taliks, they present problems to design infrastructures, especially at marine deposit such as isostatic uplifting area or coastal active area. In northern Alaska and Chukotka, food in some of the native ice cellars located in easily excavated cryopeg layers was attributed to the self-maintained moisture from the cryopeg.

Freshwater taliks ("thaw bulb") are often located beneath the deeper water bodies (>2 m), including river channels. Brewer (1958a, b) reported that the talik zone beneath a deep lake at Barrow reached a depth of 60 m. Freshwater taliks are also present under many of the closed-system pingos and drained lakes in an area which extends from North Slope of Alaska to Mackenzie Delta in Canada e.g., Mackay 1997; Parameswaran and Mackay 1996). These freshwater taliks have a relatively short life, in contrast to brine layers in regions of thick continuous permafrost. Brown (1969) observed a high saline zone several meters below sea level at Utqiagvik (formerly Barrow) Alaska. Collett and Bird (1993) compiled drill logs from an oil field in the Prudhoe Bay/Kuparuk River area, where brine layers occur widely and extend offshore at depths of 50 m to more than 250 m. The rule governing the physical properties of freshwater taliks and brine layers is similar to liquid phase of H_2O, which has both the low resistivity and high dielectric constant. However, the thermal stability of a talik and a cryopeg layer has quite different regimes. A freshwater talik is only stable when the upper boundary temperature is above 0 °C or water is moving in the ground. On the other hand, the cryopeg layer is more stable for the long period of time without any conditions in the permafrost, because the cryopeg layer stays at similar temperature of surround permafrost.

There are many ways to develop a freshwater talik in the Arctic. Simply, a talik could establish by groundwater moving in an aquifer or in the bottom of a lake or, a river. Refreezing of the talik is an important process of pingo development. A pingo is a significant landform on the flat tundra of the Arctic. In this area of low vegetation and little elevation difference, a pingo is commonly the most prominent feature against an otherwise unbroken horizon. Pingos provide a target destination for polar travelers and a dry place in the surrounding wet tundra. Ancient hunters used pingos when hiding from animals. The artifacts and archeological sites found at pingos on the Alaska North Slope date to about 6000 years B.P. (Lobdell 1986). Porsild (1938) described and classified all mounds in North America numerous mounds from the Kotzebue Region to the Mackenzie Delta. As the first person to use the term pingo, Porsild described what most likely are closed-system (hydrostatic) pingos. Pingos are ice-cored mounds, a definition used worldwide in permafrost regions. Each country or region with pingos has used its own name for this landform. Some studies before Porsild (1938) referred pingos as "heaved mounds" and "hydrolaccoliths" (e.g., de K Leffingwell 1919). The term pingo derives from the word pinga in the Inuit (Canadian Arctic) language or ping in the Inupiat (Alaskan Arctic) language. In Russia, the equivalent term for pingo, used by Yakuts, is bulgunniakh. Many pingos in Greenland were noted as mud volcanoes on maps, though actual mud volcanoes are rare, and possibly nonexistent in Greenland. Though pingos are still commonly referred as mud volcanoes in Greenland, maps published there since the 1980s use the word pingo. In Svalbard, pingo is a well-known and frequently used term. Orvin (1944) studied springs and mounds, calling them kildehaug, a term used only at that time.

Pingos have different origins, based on source of water and types of developed groundwater pressure. Classically, pingos are divided into two types, depending on the mechanism that pressurizes the groundwater, therefore pingos are either open-system (hydraulic) or closed-system (hydrostatic). Open-system pingos are formed by artesian pressure (Müller 1959; Holmes et al. 1968; French 1996). Water is supplied under pressure from surrounding higher terrains, it then flows beneath the surrounding impermeable permafrost, and finally discharges toward the ground surface through a talik at the base of a hill or in the valley bottom. In the discontinuous permafrost zone (e.g., warmer permafrost temperature), hydraulic conductivity is higher through taliks above, below, or within the permafrost. Thus, most open-system pingos occur mainly in the discontinuous permafrost zone near areas of marked relief, which supplies the hydraulic gradient (potential). Discontinuous permafrost provides the opportunity for upwelling (open talik environments). Artesian groundwater environments cause repeated injections of water into the weakest portion of the permafrost, leading to development of massive ice at the upper part of the permafrost and formation of the pingo's core. This type of pingo is commonly found in valleys near terminal moraines, at the toe of alluvial fans near river bottoms, in highly fractured or faulted uplands, and on rebounding marine terraces (Yoshikawa and Harada 1995).

In contrast to open-system (hydraulic) pingos, closed-system (hydrostatic) pingos occur in regions of continuous permafrost (Mackay 1979, 1998), where there is an impermeable layer at depth (closed talik environments). Most closed-system pingos form in drained thaw lakes or former streambeds, where the saturated lake basin sediments are exposed to the atmosphere. This exposure causes saturated sediments to freeze, forming new impermeable permafrost. However, residual ponds within the lakes retard permafrost growth, resulting in locally thinner permafrost and less resistance to deformation. As the permafrost aggrades, the pore water is expelled ahead of the freezing front. Blockage of this pore water flow by the continuous permafrost at depth redirects the flow inward from the basin edges. Injection of expelled water into the overlying permafrost, followed by freezing of the injected water, causes progressive doming. This doming, commonly situated beneath the least permafrost aggradation, produces the massive ice that forms the core of the pingo.

For both types of pingos, the 9% expansion due to phase change from water to ice is insufficient to account for size; groundwater pressure is required to supply the volume of water to a pingo ice core. Both open- and closed-system pingos may undergo vertical pulsing (periods of uplift) because of the injection of groundwater to a sub-pingo water lens (Mackay 1998; Yoshikawa 2008). To summarize, closed-system pingos characteristically (1) contain a core of injection ice (frozen from bulk water), (2) occur on lake sediments but also on previously unfrozen sediments of abandoned stream channels, and (3) are caused by the expulsion of pore water when saturated coarse- or fine-grained lake sediments freeze (Mackay 1979).

There was a unique artesian sub-permafrost groundwater leakage problem in Fairbanks Alaska next to an open-system pingo. Leakage of groundwater on the outside of residential well casings is a common problem. An impermeable permafrost layer produces unique artesian conditions at base of a hill slope (permafrost acts as an aquitard). When a well is drilled through the permafrost, the water flowing in the well induces thawing outside the well casing. To stop the flow, the unfrozen soil around the casing must be refrozen or the hydraulic conductivity reduced (by grouting). Various techniques have been used for wells around Fairbanks, Alaska. In one case, liquid nitrogen was used to stop the flow around an artesian well (Fig. 14.1). In November 2005, liquid nitrogen was injected into the well. By February 2006 the area around the well was completely refrozen, stopping the groundwater discharge. We had the opportunity to monitor sub-permafrost groundwater pressures during and after this event. After the injection, well leakage was eventually stopped and the groundwater pressures to recover. Within a few weeks, the pressure returned to the original level (Fig. 14.1). In this setting, frozen silt layer is 27 m thick and overlies 15 m of creek gravel. The permafrost is 43 m thick. The hydraulic pressures of the nearby pingos and alluvial deposits are high (>68 kPa) measured by surrounding artesian well water from sub-permafrost aquifer, thus creating artesian conditions. Most of the homeowners have installed groundwater wells for domestic usage. One wants the water inside the well casing to remain unfrozen, but the soils around the casing to remain frozen, a challenging task. The discontinuous permafrost temperatures around Fairbanks are very warm (>–1 °C) and sub-permafrost groundwater is just a few degrees above the freezing point of water. In some cases, the ensuing discharge from the wells continues to flow through the winter, thus creating large masses of aufeis (icings), some of which have destroyed houses and covered roads.

Surface disturbances have also produced talik formations throughout time. A typical natural ground surface disturbances is wildfire. Wildfire effects on frozen ground include short-term increase in soil moisture content, followed by a long-term decrease in soil moisture content, warmer soils, and increased active layer depth. The short-term increase in soil moisture is caused by removal of vegetation, which causes decreased evapotranspiration. The thermal conductivity of the active layer soils is greatly increased by this increase in soil moisture content

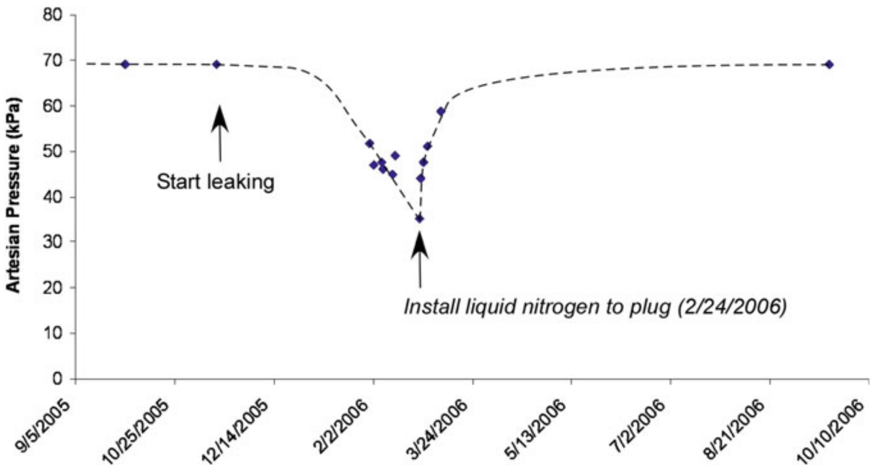

Fig. 14.1 Time series of an artesian pressure of groundwater. Liquid nitrogen was used to stop the flow leaking ground water around an artesian well

causing warmer soils than in adjacent unburned areas (also due to reduction in surface albedo and lack of surface organic material). However, the severity of the fire in part determines the amount of effectiveness of the increased soil thermal conductivity. There appears to be a threshold of the remaining organic material that determines the influence of wildfire on the frozen ground. This threshold is a function of the thickness of organic layer, thermal conductivity, and thawing index of the ground surface.

Similar disturbance to the permafrost can also occur because of urban development and agriculture. We have one of the longest disturbance studies at Fairbanks, Alaska. Comparing current results from this investigation with Linell (1973), it is apparent that in both of the plots where vegetation was removed, the permafrost retreated downward for 26 years, with the permafrost table eventually stabilizing at the partially disturbed site. This is likely due to the reestablishment of a boreal forest at the site within 25 years. At the site where all the surface vegetation and organic material were removed the permafrost table has continued to migrate downward for the past 35 years (Fig. 14.2). Vegetation at the site has continued to evolve and is currently migrating from a shrub-birch-willow forest to one with a higher density of spruce trees and moss (Douglas et al. 2012).

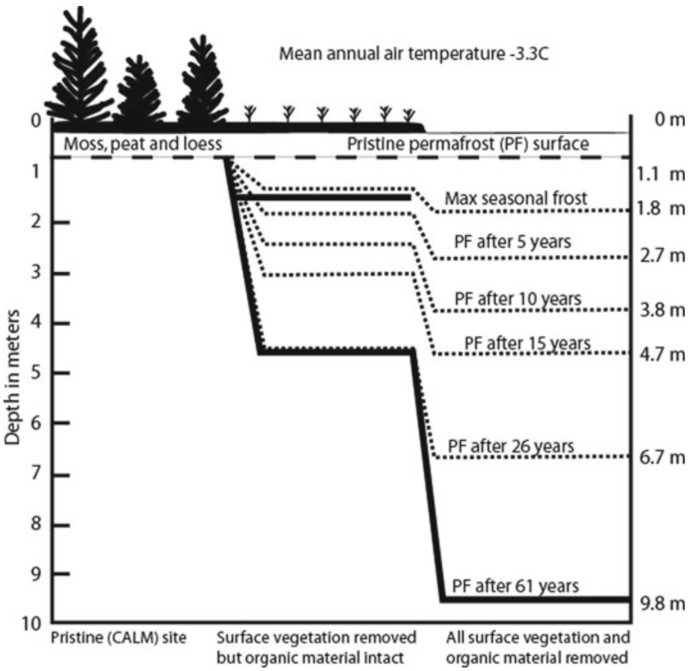

Fig. 14.2 61 years of surface vegetation and organic material removal and degradation of permafrost. A cross section based on Linell 1973. Vegetation removal from the two disturbed sites occurred in 1946. The bold line denotes the measurements collected in the fall of 2007 representing 61 years since disturbance (Douglas et al. 2012)

14.2 Historical Background

A majority of the world's permafrost is found in Eurasia including Siberia, Far Eastern Russia, Scandinavia, Arctic Islands, Mongolia, Tibet, Central Asia, Northeastern China, and Japan. Siberia is definitely the major location of permafrost with thickness from a few meters to over 1200 m. Also, Siberia has the oldest permafrost studies which originated in the nineteenth century. It means they collected temperature records since the Little Ice Age, which is very rare in the world. Permafrost temperature profile at the Shergin well (Fig. 14.3) shows little change since 1835 at a well in the middle of Yakutsk, after the town had an expanding the city limit during the last 180 years.

The main indicators of the geocryological conditions are the depth of active layer, permafrost temperature and the thickness of the frozen strata. For Siberia, the average annual temperature varies from +5 to –15 °C. The depth of active layer varies, depending on the conditions of heat exchange and climatic conditions, from 0.2 to 3.5 m, reaching 6.0 meters in mountain areas and foothills. Continuous

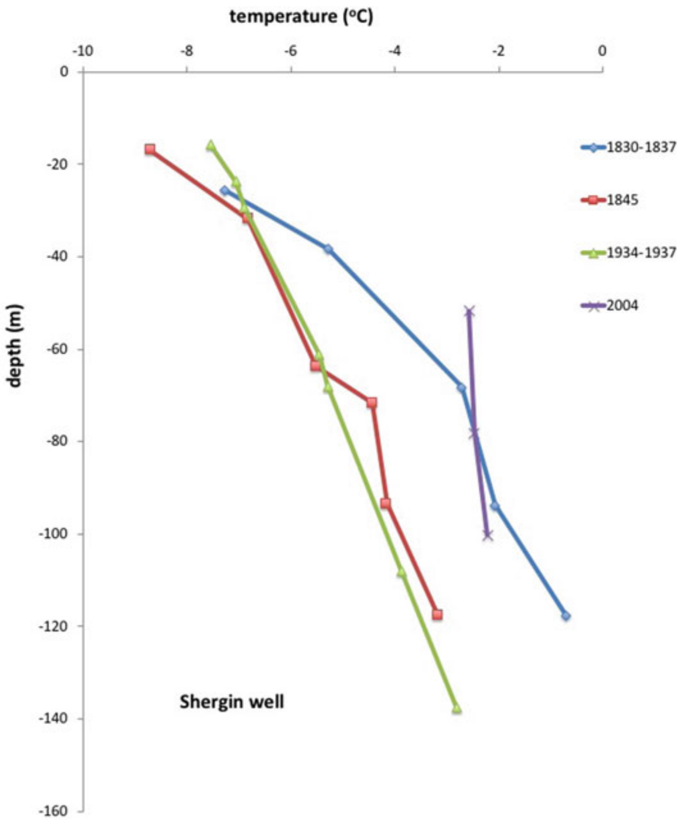

Fig. 14.3 Temperature profile in the Shergin mine, dug in Yakutsk in 1835. 1830–1837—measurements by Shergin; 1845—the measurement of Middendorf; 1934–1937—measurements of the Zatsepins; 2004—modern measurements

distribution of permafrost across the area is observed in the northern regions. But even there, under large rivers such as Lena, Yenisei, Kolyma, lakes, and in tectonic fractured zones with increased circulation of groundwater, there are zones that are absent of permafrost. Typically very complicated talik formations in such areas are difficult to predict. The dynamics of the interactions between permafrost and climate change are not easy to quantify and predict in different regions of the world. It depends on the features of atmospheric circulation processes, the conditions of heat exchange at the ground surface and the composition of soils. Over a 30-year-period in Siberia, the average annual air temperature everywhere tends to increase by 0.3 and 0.6 °C/10 years. The ice content of the permafrost is quite high in Siberia (called Yedoma deposits), including some in Alaska and Western Canada. The origin of the Yedoma deposit is syngenetic ice wedge network. These ice wedges grew upward following deposition of the eolian sediments. Most of the ice was formed during late Pleistocene (30–40 k. y, BP). Huge outcrops are seen along the East Siberian Sea coastline and is experiencing rapidly erosion. In addition to northern Yakutia, some Yedoma deposits exist in central Yakutia. However, this high volume of the Yedoma ice does not significantly contribute to the local hydrology, even when it melts dramatically. There was a catastrophic Yedoma ice melting event in the early Holocene. During that epoch, most of the upper part of permafrost thawed and produced almost all of today's thermokarst lakes and alases. The volume of the annual thaw of ice-rich permafrost is very minimal; watershed hydrology is impacted more by the annual formation (from baseflow) and thaw of aufeis.

Other than the cold stable permafrost in Eastern Siberia, most of the permafrost areas have temperatures near the thawing point and are unstable. These warm permafrost areas generate more taliks, both inter-, subpermafrost. As a result, icings and/or open-system pingos are more common, especially in Mongolia and Tibet. These intra- permafrost aquifers can result in pingos and aufeis formations, but also connect to the groundwater system through taliks under rivers and lakes, as well as terrain at higher elevations. These hydrocryological structures (pingos, taliks, icings, etc.) developed mainly from early to middle Holocene. Also, it still happens that new springs and associated icings can develop after strong seismic activity in the southern fringes of permafrost, sporadic and/or isolated permafrost can be found (some as shallow as 10 m) beyond where continuous and discontinuous permafrost is found. These permafrost areas have a unique, thicker peat layer that shields the shallow permafrost from thawing, it is common to find this in places such as northern Scandinavia, Kola peninsula, western Siberia, Kamchatka, Sakhalin, Hokkaido, and Northeastern China. Typically, segregation ice developed in the peat layer during freezing process, thus forming palsas. Annual mean ground surface temperature can sometimes exceed +4 °C in these palsas, but the permafrost can have a stabilizing. Dried peat layer has very low thermal conductivity compared with wet frozen peat layer. These thermal property differences between the seasons makes for a strong thermal offset that results in stable permafrost in such relatively warm areas.

Permafrost is widely distributed in the high latitudes of North America (mostly Alaska in United States). However, geological and permafrost chronological distribution is quite different between eastern and western parts of North America. Much of Alaska and western Canada has thicker sediments with ice-rich permafrost; sometimes buried glacier ice also exist, mainly in Northwestern Alaska, Brooks Range, Yukon Territory, and Northwest Territory. These relict glacier ice bodies have produced some huge retrogressive thaw slumps in these areas and resulted in increased suspended sediment in river/coastal waters. It is not only buried glacier ice that is found in these areas, but also pingos and both types of ice wedges (syngenetic and epigenetic). Unlike Western North America, Eastern Canada and Greenland are covered by very old igneous and high grade metamorphic bedrocks. Also, this area was covered by an ice sheet during the last glacial maximum (LGM) that prevented permafrost expansion at depth. As results, most of the bedrock area had less ice contents, except in active fluvial terrain and wetlands.

Inter-permafrost hydrological activity is very active in most of the Alaska and Canada except the area covered by Canadian Shield. Many springs, icings, and open-, closed- system pingos are located in these areas. Glacial isostatic rebounding is another important process here. Very dynamic and formed/emerged beach lines established complicated permafrost hydrological interactions in Eastern Canada and Greenland. These areas have also considerable marine clay deposits that harbor cryopegs.

Most of the western Canada and interior Alaskan boreal forests are underlain by discontinuous permafrost. The discontinuous character of permafrost distribution creates a complex interrelationship between permafrost and vegetation in the boreal forest. Especially the type of forest floor varies from litters, lichens, feather mosses to Sphagnums. Permafrost thicknesses vary in this area from 0 to 150–200 m (Ferrians 1965; Brown et al. 1998). Permafrost temperatures are close to 0 °C, thus the permafrost is more susceptible to surface disturbances such as wildfire. Permafrost temperatures in the boreal forest are 0 to –4 °C and typically warmer than –2 °C. Other natural factors that influence permafrost temperature regime include the thickness, thermal properties, and duration of the snow cover (Brown and Pewe 1973; Romanovsky and Osterkamp 1995). The mean annual ground surface temperatures usually are 3–6 °C warmer than the mean annual air temperatures. As a result of relatively warm air temperatures and the effect of snow cover in reducing winter heat loss from soil, the mean annual ground surface temperatures in the boreal forest often exceed 0 °C and can be as high as 4 °C. Isolated permafrost was found to have a strong thermal offset in Bristol Bay area of Alaska, Anchorage Alaska and the Whitehorse area of Canada. Based on current borehole information, one of the coldest permafrost areas in the arctic and subarctic is the measured –27.6 °C near the summit of Denali (station elevation ca.5700 m). However, for low elevations, Alert (northern most end of the Canadian archipelago) was measured at –14 °C. All of Eurasian permafrost temperatures were warmer than Alert and Denali.

14.3 Research Highlights

14.3.1 Aufeis and Its Contribution to Base Flow

Significant amounts of aufeis are commonly observed in the permafrost regions of the northern hemisphere such as Alaska (Sloan et al. 1976; Slaughter 1982; Carey 1973), Arctic Canada (Pollard 2005), Yukon Canada (Harris et al. 1983), Svalbard (Liestøl 1977), Greenland (Yde and Knudsen 2005), Siberia (Alekseev and Tolstikhin 1973; Sokolov 1973), Mongolia (Froehlich and Slupik 1982), and Tibet (Zhou et al. 2000). The term “aufeis” is of German origin and roughly translates to “on or upon ice.” Icing (English) and naled (Russian) are also widely accepted terms in the scientific community. However, aufeis is the term more frequently used in Alaska, and is therefore used in this text. The history of aufeis research dates back more than 150 years. Wrangel (1841) observed aufeis along the northern shore of Siberia. During the mid-nineteenth century, some Siberians explored and wrote about mounds and icing blisters (e.g., Middendorf 1861; Maydel 1896). Podyakonov collected detailed observations of river icing formations to develop the "Padyakonov formula" (Podyakonov 1903). This is an empirical equation that tries to identify those characteristics that enhance or reduce aufeis formation. Many researchers (e.g., Sumgin 1927) have examined this formula and modified several inherent limitations.

Sumgin was one of the first Russian scientists to study aufeis, and used the word "naled" to describe them. Parkhomenko (1932) studied ice formation from sub-permafrost and intrapermafrost groundwater in more detail. Tolstikhin pointed out that the river aufeis often developed in the areas where sub-permafrost water springs discharged into the river bottom. It is also common for the springs to be in the near vicinity of the stream. The volume of aufeis associated with the direct outflow of a source is determined by the following equation (Tolstikhin 1941):

$$V = Q * t * k * A$$

where:

V = aufeis volume (m^3),

Q = outflow of source (m^3/h),

t = time required for development of aufeis (hours),

k = coefficient of aufeis maturing, and

A = empirical coefficient depending on runoff conditions, evaporation, precipitation, and condensation at the ice body surface.

Tolstikhin (1963) empirically calculated aufeis runoff as a function of specific aufeis surface. The specific aufeis surface is based on following expression:

$$Fu\,F = Q$$

where Fu = specific aufeis surface, i.e., the gain in aufeis area per unit of outflow (m/s),

F = aufeis area (m^2), and Q = spring discharge (m^3/s).

Hall and Roswell (1981) tried to determine the coefficients and estimate groundwater discharge rates in the Brooks Range on the basis of Sokolov's formulation at five aufeis sites using 1972–1979 Landsat imagery and obtained:

$$V = 0.96\,F$$

Åkerman (1980) also added "glacier icing" and "pingo icing," which are located downstream of the glacier's terminal moraine and associated with open-system pingo formation, respectively. These are both common locations of groundwater emergence at the ground surface in permafrost regions. These aufeis categories are also mainly distinguished by having different sources of water. As expected, spring aufeis forms on or just downstream of a groundwater spring. An ice mound is frequently observed at the spring in early or mid-summer. River aufeis is widely developed in valley bottoms, usually some distance downstream of a spring source. Ground aufeis forms when active layer water (suprapermafrost groundwater) seeps to the surface.

The chemical composition of the source water contrasts strongly between suprapermafrost, intrapermafrost, and sub-permafrost groundwater sources. Ground aufeis is highly concentrated in dissolved organic nitrogen (DON) and dissolved organic carbon (DOC). It is possible to determine the source of the water for each type of aufeis by measuring the dissolved inorganic carbon (DIC) and DOC ratio (DIC/DOC) (Yoshikawa et al. 1999). Spring aufeis has higher DIC and lower DOC as compared to other types of aufeis.

Ground aufeis also has a very high DOC content and commonly has a dark brown color due to the source water flowing through a layer of organic material. The characteristics of river aufeis lie between those of ground and spring aufeis. Romanovskii (1983) pointed out that the aufeis formation is not only affected by the sources of water, but also controlled by the permafrost conditions (e.g., continuous, discontinuous, or sporadic). Groundwater springs and associated aufeis are not uniformly distributed throughout the North Slope of Alaska. More than 30,000 l/s. of spring water discharge along the eastern part of the foothills of the Brooks Range (Kane et al. 2014; Childers et al. 1977). These springs with steady flow all year around result in large areas covered with aufeis every winter (Harden et al. 1977). Aufeis is second only to snow cover as the biggest temporary surface storage reservoir of fresh water during the winter period in unglaciated basins. Kane and Slaughter (1973a, b) and Slaughter (1982) suggested that up to 40% of winter streamflow (4% of the volume of annual runoff) may be stored as aufeis in a small watershed in interior Alaska. The reflection and emittance of electromagnetic radiation varies between frozen and unfrozen water. Many different remote sensing systems utilize this distinction to discriminate liquid water and ice. This is a common method for studying and characterizing aufeis deposition over wide areas (e.g., Landsat (Dean 1983; Harden et al. 1977; Hall and Roswell 1981) and SAR (Li et al. 1997)). Li et al. (1997) estimated aufeis formations using complex SAR data (interferometric SAR) at Ivishak River, Alaska. Accurate regional assessments of aufeis location and extent are now available because of new remote sensing techniques that have been recently developed using a combination of multiple wave bands as well as finer temporal resolution.

Seasonal characteristics of aufeis development and thawing were monitored at the aufeis field between 2004 and 2005 in the Kuparuk River basin north of Toolik Lake (Yoshikawa et al. 2007). The Kuparuk aufeis field is driven by a large spring that emerges about 2 km upstream of the aufeis. The discharge is around 800–1000 l/s. The water temperature is slightly below 0.0 °C. Analysis of SAR imagery from 1996 to 2006 indicated that 3–18% of the maximum aufeis accumulation remained at the end of the summer. The end-of-summer ice area seems to have increased between 1997 and 2005. The soil temperatures beneath the remnant aufeis after summer melt are never above 0 °C. Where remnant summer aufeis persists, stable or aggrading permafrost is found. This will have a positive feedback

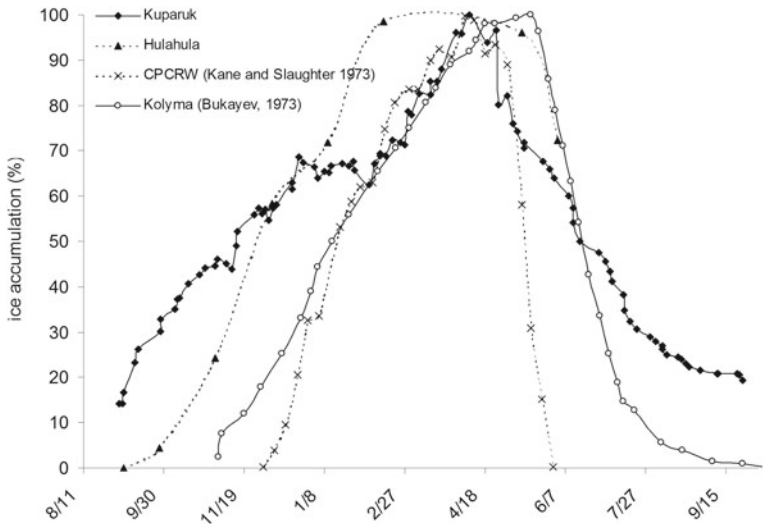

Fig. 14.4 indicates that aufeis in the Brooks Range starts to develop around October and continues to grow through early May; SAR image analysis for the Kuparuk River (2004–2005) and Hulahula River (2002–2003) throughout the winter were used for mapping

effect on the remaining summer aufeis and will also impact the geometry of the groundwater pathway above the permafrost.

Figure 14.4 Indicates that aufeis in the Brooks Range starts to develop around October and continues to grow through early May; SAR image analysis for the Kuparuk River (2004–2005) and Hulahula River (2002–2003) throughout the winter were used for mapping. Also Fig. 14.4 shows for comparison both the growth and decay of aufeis at Caribou-Poker Creeks Research Watershed (CPCRW) (Kane and Slaughter 1973a, b) representing warm discontinuous permafrost region and Kolyma River (Bukayev 1973) representing a Siberian continuous permafrost region. The North Slope of Alaska aufeis formations increase dramatically in late winter and the development of aufeis starts earlier in colder permafrost regions in response to earlier and longer winters than in the discontinuous permafrost regions.

Over the winter season about 50% of the aufeis has already formed on the North Slope of Alaska before December, more than 1 month ahead of time relative to the warmer discontinuous permafrost region to the south. The time series of the SAR images reveal the formation of the aufeis development. Aufeis fills the stream and river channels in early winter (stage 1). Aufeis develops a thicker and smoother surface during the middle of the winter (stage 2). Once a massive and smooth ice body has developed, aufeis starts to grow thicker and expand downstream in late winter (stage 3). Continued overflow water spreads widely into the floodplain and reaches farther downstream (often without freezing at the surface in late winter).

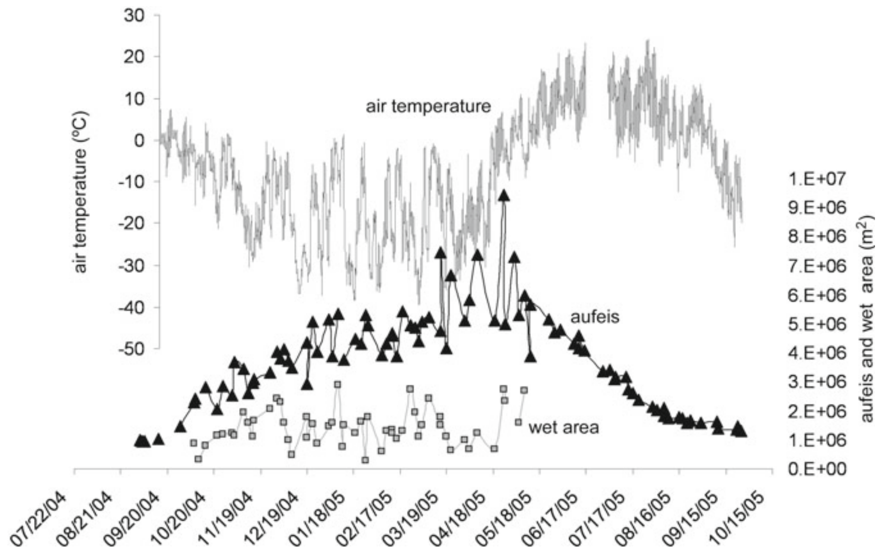

Fig. 14.5 Air temperature at the Upper Kuparuk River meteorological station near the Dalton Highway (about 30 km south of aufeis field) and overflow (wet surface) intensity during 2004–2005

Figure 14.5 shows the air temperature at the Upper Kuparuk River meteorological station near the Dalton Highway (about 30 km south of aufeis field) with overflow (wet surface) intensity during 2004–2005. The wet area was estimated using SAR imagery, which helps to understand aufeis formation processes and timing of growth. We cannot estimate the wet area during thawing season (before 20 October 2004 and after 18 May 2005). The timing of the aufeis formation corresponds with fluctuations of the air temperature. Overflow water on top of the aufeis principally occurred when large air temperature fluctuations occurred within a 2 day period.

Kane (1981) described a similar phenomenon in aufeis activity near Fairbanks, Alaska, through the use of piezometers. The cause of the overflow discharge is related to the increase of hydrostatic head during a warm period following a colder period. He observed maximum heads during the warmest periods during the winter and the lowest heads during extended cold periods. The thermal stability of the inner unfrozen channels in the aufeis appears to be an important parameter of aufeis development.

Lifshits et al. (1966) defined an aufeis melting coefficient (γ, mm/°C d) as follows:

$$\gamma = H/\Sigma t$$

where H = amount of melting (mm) and Σt = sum of average daily positive air temperature (thawing index). The average total amount of melting per year during the period of 1996–2005 was 14,410,000 m^3/a, which requires a melting coefficient of 2.19 mm/°C d at the Kuparuk aufeis 1.90 mm/°C d in Sadlerochit spring aufeis and 3.62 mm/°C d in upper Lena River (Lifshits et al. 1966) aufeis. Thawing index is an important indicator of the rate of aufeis degradation. Colder regions have lower melting coefficients than the warmer discontinuous permafrost regions. The proportion of aufeis runoff (volume of aufeis) to total annual groundwater discharge is 27–30% for the Kuparuk River site. During the summer months, the Kuparuk aufeis releases water to the stream, especially in June and July. About 16 million m^3 of base flow from spring discharge went into storage as aufeis in the winter of 1999/2000. Approximately 1.1 million m^3 of aufeis carried over into the winter of 2000/2001 and approximately 15.9 million m^3 of new ice formed during the following winter. Aufeis activity is a sensitive indicator of winter base flow (spring discharge) and spring water temperatures. Analysis of periodic aerial photography collected over the past 50 years for the Hulahula, Sadlerochit, and Kongakat Rivers in Alaska indicate that the aufeis fields have not dramatically changed in either volume or extent. Throughout the Landsat era, the Kongakat River aufeis field has kept a constant volume as in the past (Harden et al. 1977). The Kongakat River aufeis has one of the longest records of observations by explorers (e.g., Franklin 1828) because it is one of the biggest ice fields and located close to the coast and therefore easy to observe from the Arctic Ocean. Norwegian explorer, Roald Amundsen described this aufeis field as a glacier in 1906 during his first voyage (Amundsen 1908).

14.3.2 Ground Ice Volume/Change and Its Impact on Regional Hydrology

Active ice wedge polygons may be observed near the southern boundary of the permafrost in areas such as Goldstream Creek valley in Fairbanks Alaska. In general, the frost contraction cracking process is inactive in the Fairbanks area under the contemporary climate. Many of these ice wedges developed 32,000–39,000 yBP (Marine Isotope Stages—3). However, Holocene (current) ice wedges are present in this area. Frost contraction cracking still occurs occasionally about every 10 years or so, particularly, during low-snow, severe winters in areas with substantial micro topography such as well-developed tussocks. Tussock tundra is a common vegetation in interior, western and northern Alaska. Tussocks develop as earth hummock type mounds (ca. 50 cm diameter) with relatively deep (20–50 cm) air-filled annular spaces around the tussock (sometimes filling with standing water). This micro topography produces conditions that are very difficult for people to traverse. Tussock tundra also experiences colder thermal conditions due to non-conductive heat transfer. Cooling is enhanced by convective heat transfer during winter months, evapotranspiration during the summer, and blocking of direct solar radiation by the rough surface tussock vegetation. This rough surface topography helps keep the ground cooling and is similar to the effects of rock glaciers or block fields.

Most areas in the southern boundary (Northern Hemisphere) of the mountain permafrost are occupied by rock glaciers or frozen block slopes. Rock glaciers or block slopes have the significant pore space available for convective heat transfer during winter periods. Woodcock (1974) reported 10 m of thickness of the permafrost at Mauna-Kea, Hawaii, despite a mean annual ground surface temperature at the summit of Mauna-Kea of +1.2 °C. Permafrost or massive ice body has been observed in many warm (positive ground surface temperature) locations due simply to the site-specific heat transfer process that is enhanced by convection during the winter but diminished during the warmer season. Goering and Kumar (1996) designed a road-bed that promotes winter-time convection in open graded embankments for cooling of the underlying, unstable permafrost. The road embankments are cooler than the surround ground during winter periods, driven by convective heat transfer as the cold, dense air circulates in the large pore spaces. During the summer, the surface is warmer and convective heat transfer is minimal as the warmer, less-dense air mass in the pore space does not sink. This type of structure is commonly seen in cold region engineering designs today such as the Tibetian-Qinghai railway. Non-conductive heat transfer is a very important process influencing the stability of permafrost in some unique settings. Also, these cooling processes increase the chance of secondary intrusive ground ice formations (such as ice wedges or rock glaciers) which are likely to remain because of topographical conditions in warm permafrost areas.

Permafrost stability in discontinuous permafrost regions is strongly controlled by soil type and the physical and thermal properties of the surface active layer. In general, the organic rich (peat or moss) layer increases the stability of permafrost, perhaps to the extent where the mean annual ground surface temperature may reach a positive temperature, due to the effect of the thermal offset (Gold et al. 1972). The thermal offset describes the process where more heat may escape from the active layer in the winter than enters the soil in the summer, because of the difference between frozen and unfrozen thermal conductivity of the surface soils.

14.3.3 Thermokarst and Open Talik Lake

It is important to understand the role that permafrost degradation plays in affecting the surface water balance and subsurface thermal dynamics. Soil moisture storage in the active layer is a key variable in understanding most ecological process interactions and atmospheric/terrestrial linkages in arctic regions (Boike et al. 1998; Romanovsky et al. 2002). The primary control on local hydrological processes in northern regions is dictated by the presence or absence of permafrost, but it is also influenced by the thickness of the active layer and the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in areal extent, the interaction of surface and sub-permafrost groundwater processes becomes more important (Woo 1986). The inability of soil moisture to infiltrate to deeper sub-permafrost groundwater zones due to ice-rich permafrost produces and maintains near surface wet soils in arctic regions. However, in the slightly warmer regions of the subarctic, the permafrost is thinner and discontinuous. In permafrost-free areas, surface soils can be quite dry as infiltration is not restricted, thus impacting ecosystem dynamics, fire frequency, and latent and sensible heat fluxes. Other hydrologic processes impacted by degrading permafrost include increased winter stream flows (Yang et al. 2002), decreased summer peak flows (Bolton et al. 2000), changes in stream water chemistry (Petrone et al. 2000), and other fluvial geomorphological processes (McNamara et al. 1999). Hydrologic changes or changes that influence hydrologic processes documented among study sites include drying of thermokarst ponds, increased active layer thickness, increasing importance of groundwater in the local water balance and differences in the surface energy balance (Carr 2003).

In response to some imposed disturbance, such as a tundra fire or climatic warming, ice-rich permafrost may differentially thaw, creating irregular surface topography. Depressions forming on the surface soon form ponds, accelerating subsurface thaw through lower albedo and additional heat advected into the pond through runoff. In time, a thaw bulb or talik (a layer of unfrozen soil above the permafrost and below the pond) may form as the depth of water becomes greater than the amount that can refreeze during the winter. If the talik grows to a size that completely penetrates the underlying soil or connects to a subsurface layer that allows continued drainage, the pond may then begin to drain. Some ponds on the Seward Peninsula, Alaska are now exhibiting this behavior (Yoshikawa and Hinzman 2003). An ice wedge polygonal network exists outside of these ponds; both frost contraction and ice wedge cracking are unlikely to occur in the current warm climatic conditions (Lachenbruch 1962). This ice wedge terrain also has highly developed thermokarst in response to the recent warmer climate. Most of these ponds had drained previously (via bank rupture). Remaining ponds continue shrinking in surface area. The surrounding marsh areas are also draining with newly generated Palsas providing opportunities for spruce invasion.

Thermokarst topography forms as ice-rich permafrost thaws, either naturally or anthropogenically, and the ground surface subsides into the resulting thawed voids (Brown and Grave 1979a, b; Hinzman et al. 1997). The important processes involved in thermokarsting include thaw, ponding, surface and subsurface drainage, surface subsidence, and related erosion. These processes are capable of rapid and extensive modification of the landscape; predicting, preventing or controlling thermokarsting is a major challenge for northern development (Lawson 1986). Many thermokarst ponds and depressions have been observed across interior Alaska and Canada in regions of discontinuous permafrost (Jorgensen et al. 2001; Osterkamp et al. 2000; Burn and Smith 1990). Osterkamp and Romanovsky (1999) observed permafrost temperatures warming by 1.5 °C since the 1980s in interior Alaska. Permafrost temperatures in boreholes displayed a 2–4 °C increase over the last 50–100 years on the North Slope of Alaska (Lachenbruch and Marshall 1986). Thermokarst and permafrost degradation is not confined to Alaska and Canada. It is occurring worldwide, with thermokarst development particularly widespread at the southern limit of the discontinuous permafrost zone in countries such as Mongolia (Sharkuu 1998), China (Ding 1998), and Russia (Pavlov 1994; Czudek and Demek

1970). The dynamics of thermokarst lakes in areas of thick permafrost in the Arctic are probably associated with a changing surface water budget in response to some catastrophic event impacting near surface ice-rich materials (Sellmann et al. 1975). However, thermokarst processes are much more active in Subarctic regions of discontinuous permafrost. In the Subarctic, permafrost temperatures are frequently warmer than –1 °C and intra- or sub- permafrost water flow is often observed (Yoshikawa et al. 2002). The thermokarst water movement of the Subarctic is more complex than in the continuous permafrost regions. Ice-rich permafrost can be highly impermeable aquitard, frequently causing formation of artesian conditions (i.e., upward hydraulic gradients in low elevations). Kane and Slaughter (1973a, b) reported on the hydraulic processes associated with a lake recharged from a sub-permafrost aquifer near Fairbanks, Alaska. The thaw bulb below the pond had thawed completely through the permafrost thus enabling sub-permafrost groundwater to recharge the pond. In this case, the pond level was more stable as compared to ponds recharged only from surface runoff. Sub-permafrost groundwater is typically enriched in cations, resulting in higher electrical conductivity of pond water.

The water budget of the various pond types may be calculated as follows:

New thermokarst pond

$$Q_{\mathrm{in}} + Q_{\mathrm{thaw}} > Q_e + Q_{\mathrm{out}}$$

New thermokarst depression

$$Q_{\mathrm{in}} + Q_{\mathrm{thaw}} < Q_e + Q_{\mathrm{out}}$$

"Mature" shrinking pond

$$Q_{\mathrm{in}} + Q_{\mathrm{thaw}} < Q_e + Q_{\mathrm{out}} + Q_g$$

where Q_{in} is input water from atmosphere, active layer or surface, Q_{thaw} is water from permafrost thawing, Q_e is evaporation and evapotranspiration, Q_{out} is outflow through the surface or active layer, Q_g is drainage to sub- or intra-permafrost groundwater through a talik.

The key phenomenon of the shrinking pond water budget is quantifying the magnitude of Q_g, which is primarily dependent upon the hydraulic gradient and hydraulic conductivity, both which can vary significantly. A drier climate and/or increasing water-holding capacity of the active layer will cause a lowering of the local groundwater levels, accelerating the shrinkage of pond surface areas.

The implications of this analysis are that in regions over shallow (<30 m), warm permafrost surface ponds (~>10 m diameter) may shrink in size and newly developed small ponds may form. In addition, surface soils may become drier as the permafrost degrades. This depends upon regional hydraulic gradients (i.e., whether the region is a groundwater upwelling or down-welling zone). The same mechanisms that allow drying of the ponds may also cause soil drying with significant impacts to latent and sensible heat fluxes.

Permafrost temperatures are strongly affected by the slope and aspect in this region. In Fairbanks area, south facing slopes, in general, are absent of permafrost while permafrost is usually present on north facing slopes and in the valley bottoms. One of the mature thaw lake features is a floating organic mat covering most of the lake surface. Where more than 1 m of floating organic material exist, the surface of the lake can usually be comfortably walked over all year round. Isabella Creek bog lake near Fairbanks, Alaska, is a 7 m deep water body surrounded be permafrost with an open talik underneath. 5 l/s. of discharge from the lake was measured in 2003. This area has permafrost with artesian conditions, which means the groundwater recharges the lake water all year round. Kane and Slaughter (1973a, b) studied this lake to estimate discharge and soil hydraulic conductivity using piezometric measurements at different depths in the middle of lake. They estimated the hydraulic conductivity and discharge using Darcy's Equation. The hydraulic gradients were measured at 0.2–0.5 m/m. The rate of discharge (Q) is calculated based on the talik area (A), hydraulic conductivity (k), and hydraulic groundwater gradient ($\mathrm{d}y/\mathrm{d}z$). The piezometric pressure was measured at four different depths. A positive vertical gradient was measured all year round (1969–1970), indicating this pond was recharged via the taliks throughout the year. As mentioned above, the discharge rate was calculated by Darcy's law:

$$Q = kA\,(\mathrm{d}y/\mathrm{d}z).$$

where Q is the vertical discharge, K is the hydraulic conductivity; A is the cross-sectional area of thaw, beneath the lake, and ($\mathrm{d}y/\mathrm{d}z$) is the hydraulic gradient. Hydraulic conductivities of the materials beneath the lakes were estimated based on unfrozen silt. Field observations in 2003 further support discharge rate of earlier work (Fig. 14.6).

Understanding 3-dimensional talik geometry or groundwater aquifer would be a significant contribution to permafrost hydrology. Especially needed is improved technology necessary for detecting permafrost cryopeg pockets and aquifers in the frozen layers of warm permafrost as well as usage of the ice cellar (storage) for local people. Drilling is one approach for understanding permafrost boundaries; however, it's not suitable for all applications because of the high cost and the time commitment. Geophysical investigations will be the best method to address this problem. In this section, we discuss resistivity and dielectric constant characteristics of the liquid content of frozen ground. Resistivity experiments in the cold chamber indicate that higher resistivity values occur at temperatures colder than the freezing point of bulk water. The liquid water content of the materials also has a strong influence on resistivity (Fig. 14.7).

Kawasaki and Osterkamp (1988) developed a temperature dependent conductivity model. The apparent conductivity of this model has a strong impact on the temperature regime. The electrical resistivity of soils, sediments, and rocks to the direct electrical current is a powerful and sensitive parameter for the detection of

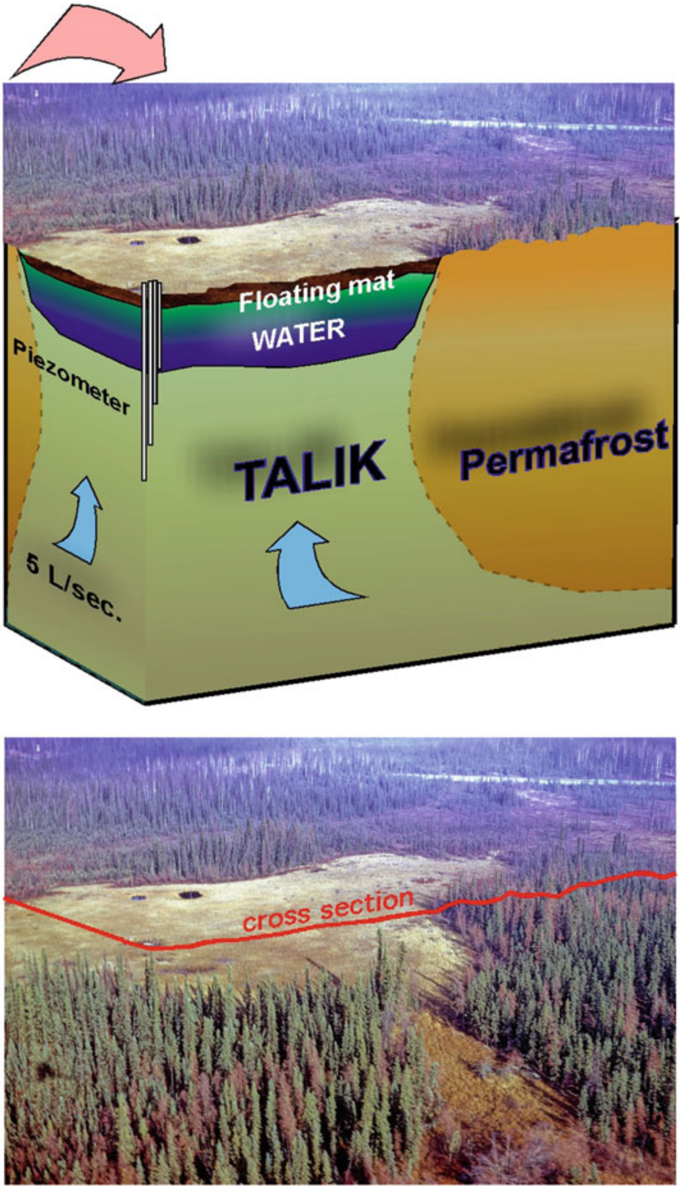

Fig. 14.6 Groundwater recharged lake. In general, bog lakes are connected through an open talik that allows recharge to the lakes. In this case, the lake water had much more minerals (higher electric conductivity) and ideal growth conditions for the floating vegetation

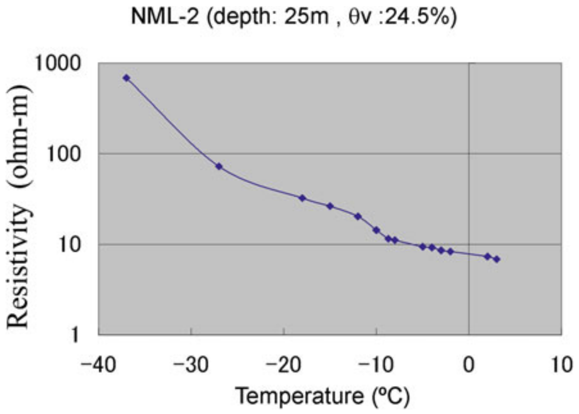

Fig. 14.7 Resistivity versus temperature regime from one of the brine samples from North Slop, Alaska. Ground temperature (–9.54 °C at 26 m depth) is the freezing point of this material in which the value increase immediately after the temperature is colder than –9.5 °C

brine layers (Duxbury et al. 2001, 2004). However, the resistivity is a less sensitive indicator of the soil type or water content under highly saline conditions. The higher frequency dielectric constant is an ideal parameter for the indication of the soil type, water content, and other physical properties. GPR surveys with the velocity analysis provide complex dielectric constants. Brine layers have an unusually high imaginary part of the dielectric constant. The imaginary part of the dielectric constant is strongly dependent on the freezing temperature. Thus, the ground temperature regime potentially controls the ionic concentration of brine waters. The annual temperature fluctuation zone in permafrost is especially important. The electric properties also change from season to season within this zone.

14.4 Permafrost Hydrology and Indigenous People

Arctic indigenous people live in the permafrost regions for many generations, their life style is well connected to the environment, including the usage of ice cellars. Ice cellars (Lednik in Russian, Bulus in Sakha, and Sigruaq in Inupiat) excavated into the permafrost layer, are a natural form of refrigeration for preserving block ice for drinking water, storing harvested food (fish, game meat, whale and livestock such as reindeer), and fermenting food. Ice cellars are traditionally used by indigenous Arctic people, such as Siberian Even, Evenk, Chukchi, Yukagir, Dolgan, and North American Inupiat, Yupik, and Inuit. Though ice cellars are widely used in permafrost regions, their structures and the purpose of their use, and the methods of maintenance are quite different among the communities due to the variations in permafrost setting. Monitoring ice cellar temperatures and recording descriptions of ice cellars are important in climate change; it is also of interest in permafrost studies and archiving traditional techniques of living with permafrost.

In Siberia, over 90% of the land is underlain by permafrost, a region where most of the indigenous people of Siberia live; and the same is true for the North American Arctic people. In the Sakha Republic Russia, not only are subsistence foods stored in ice cellars, including caribou, ducks, fish, and marine mammals, which constitute a substantial proportion of the local diet, but also ice blocks for drinking water during the summer months, especially in Central Yakutia. These "ice houses" for temporary storage of ice cubes are not only common in arctic communities, but were also widely used in Europe, United State, Canada, and other countries in early times before the use of electric refridgeration. Ice cellar infrastructure has both cultural and practical significance. Concern has been expressed recently over the impact of climate change on ice cellars and future sustainability of this life style (Kintisch 2015). This section reports the results of an ongoing education and outreach project in permafrost regions to understand and accurately report the thermal state of ice cellar temperature regimes and the surrounding permafrost environments (Yoshikawa 2013). We visited over 500 communities including Mongolian, Even, Evenk, Chukchi, Yukagir, Dolgan, Inupiat, Yupik, and Inuit, where we discussed local ice cellars with residents.

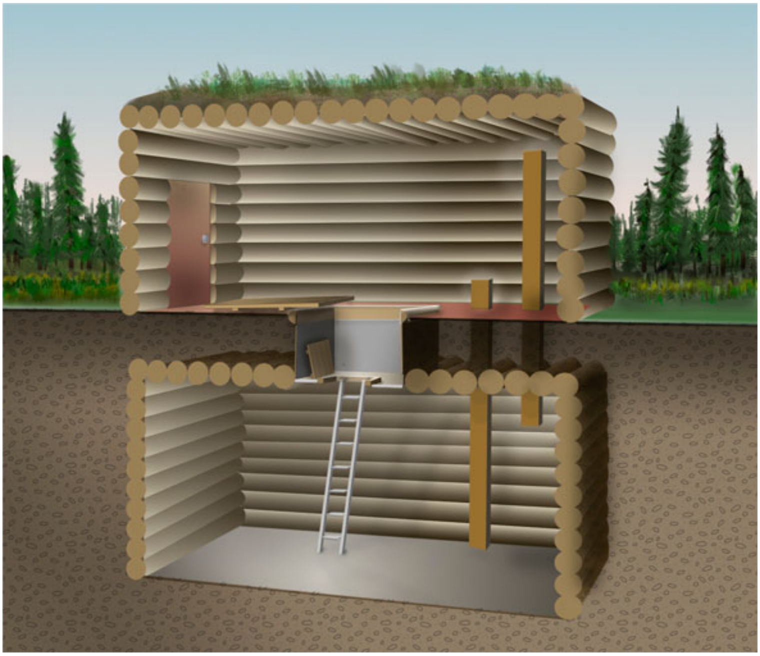

Fig. 14.8 Typical vertical cellar in central Yakutia (illustrated by M. Aoki)

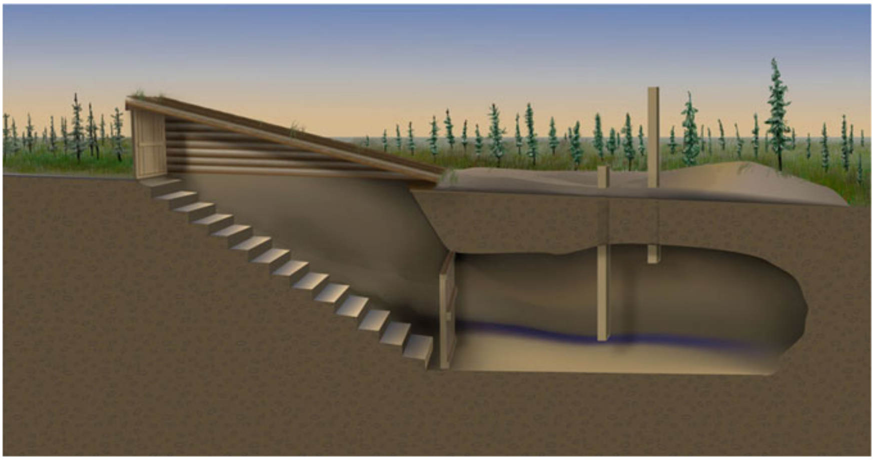

Fig. 14.9 Typical horizontal cellar in central Yakutia (illustrated by M. Aoki)

The long history of interest in permafrost and digging in frozen ground developed a variety of ice cellar types in Siberia. The Soviet Union era (about late 1950s), the Soviet government supported and encouraged the development of community-based or industrial ice cellars associated with mining activities, which resulted in unique ice cellar structures all over Siberia. Before the invention of modern day equipment, humans simply dug vertical shafts to store food in or buried food under sphagnum during summer months (M. Pogodaev, personal comm. 2016). Contemporary cellars in northern Sakha and North America regions are built primarily for personal use of one or several families (such as a fishing/whaling crews and their dependents), and typically consist of a vertical shaft that leads to a small chamber or horizontal tunnel excavated into permafrost (Fig. 14.8). These cellars vary in dimensions. The vertical shaft is 1 to 6 m deep and penetrates to a depth such that the ceiling of the chamber is below the permafrost table, which is usually less than 1 m in undisturbed areas but may be more than 2 m in villages. Older ice cellars or ones in southern permafrost areas are built primarily for personal use, and typically consist of a 15 to 20 degree declining tunnel entrance that leads to a small chamber excavated into permafrost or seasonally frozen soil (Fig. 14.9).

The depth of the chamber is 1–3 m from the ground surface. This type of cellar is similar in design to European wine and food cellars or icehouses, and provides easy access and maneuvering of what has been stored. In addition to private cellars, there are deeper, longer, and larger-capacity commercial/industrial cellars, most of which were built during the Soviet era for communities in many parts of Sakha and Chukotka regions. Many of these cellars were dug horizontally a few hundred meters, and even kilometers, into hillsides, and had railroads for managing frozen items. Though abundant, these large-capacity cellars today receive limited use;

Fig. 14.10 More than 100 years old cellar at Oymiakon, Sakha Republic where recent fall storm-flooding event and filling by ice

some have been reestablished as local museums. Recently, the media has referred to ice cellar “failures” (e.g., Kintisch 2015) or has reported that cellars no longer function reliably, because food has thawed while in storage or because an owner cannot safely access a cellar. In our visits to communities in Sakha Republic, we did not observe ice cellar failures; however, several cellars were closed due to a recent fall flooding event and filling by ice (e.g., Oymiakon, Verkhoyansk settlements, Sakha Republic Russia; see Fig. 14.10). Other than these closures, massive numbers of ice cellars were no longer manageable after the Soviet Union collapsed (e.g., Iengra, Tomtor).

Ice cellars are an efficient solution for storing large volumes of harvested ice, fish, and reindeer meat. Ice cellars require maintenance and annual cleaning; they are usually cleaned just before restocking. In northern communities, cleaning occurs in late fall for storage of marine mammals and fish during winter. In central Yakutia, ice cellars are used mostly for storage of block ice prepared in middle or late winter. Preparation involves removing all of the meat and fish from the previous year and thoroughly cleaning the cellar, including adding a layer of fresh snow.

Pond ice thickness is an important factor in the harvest and storage of block ice. The ice thickness must be 50–60 cm for cutting from a pond. Stored block ice is used mainly for drinking water during the summer months in many part of central Yakutia. Though ice storage in a permafrost cellar is ideal, the temperature inside the cellar does not have to be below zero year around; it just needs to stay cold enough to delay a rise in temperature during the warmer summer months. Underground storage this far north successfully functions because of a 4- to 5-month delayed response to rising surface temperatures. This method of cold storage is similar to how icehouses were used in the nineteenth century before electric refrigeration.

Humidity is higher in cellars, and the potential for fungus growth is always present, even in constant negative temperatures. Some cellars used for fermentation purposes take advantage of temperature and relative humidity characteristics. Fermentation is more often practiced by the Chukchi in Chukotka and by North American indigenous cultures.

Every several years, water is sprayed on the walls of vertical cellars in northern regions and horizontal industrial cellars to develop an icy surface for preventing sublimation of the permafrost. These ice-coated cellar walls minimize sublimation, keep the cellar clean, and help stabilize the walls. The annual temperature cycle of most ice cellars indicates significant cooling during winter months due to opening of doors/entrances. The maximum daily temperature observed in the ice cellars ranged from −10.4 to +2.7 °C. The temperatures in some cellars in central Yakutia rise above 0 °C. The mean annual temperatures ranged from −14.4 to −3.6 °C. In contrast, the mean annual air temperature in Yakutsk is around −9 °C. The warmest cellar temperatures were observed in fall (September to October), and typically the doors/entrances are opened to introduce cold air after November. Thus, the coldest temperatures occur during the winter months, in direct response to air temperature. The average annual temperature amplitude, i.e., the difference between the warmest and coldest temperatures, range from 9.4 to 36 °C in the cellars. This difference results from the depth of the cellar chambers, the original permafrost seasonal amplitude range, and the structure of the cellars.

Based on public media reports and our own field observations, evidence indicates several major problems related to the maintenance and use of ice cellars in North America (Klene et al. 2012; Nyland et al. 2016). Factors other than climate warming could be negatively affecting cellars, including (1) local soils known to be ice-rich and high in salinity; (2) proximity to flooding rivers or the coast; (3) influence of urban development on local hydrology; and (4) a suite of potential influences related to proximity to other types of infrastructure (Nyland et al. 2016). However, in Yakutia, the same problems as in North American cellars were not observed. Descriptions of cellar failures in Yakutia most commonly involve flooding. For large rivers in the Arctic (Lena, Yana, Indigirka, and Kolyma, etc.), flooding occurs during spring breakup due to the snowmelt following the south-to-north stream flow. Springtime floodwater in a cellar would be relatively easy to remove during the warm summer months. Though fall flooding does occur occasionally, it has the potential to be a serious problem because there may not be enough time to remove the water before the air temperatures drop, or it may be too cold to operate draining equipment. If ice fills a cellar completely, it must be abandoned for use, such as with cellars in Oymiakon and Verkhoyansk due to recent fall flooding. During the Soviet era, Kolkhoz and Sovkhoz communities operated ice cellars that were well-maintained and managed, and overseen by responsible individuals. This kind of intense maintenance helps prevent cellar failure and flooding damage in a community, but has not been economically possible since the Soviet Union era.

Air convection system in a cellar is a unique design and typically seen in Even and Evenki communities, but not in North American indigenous cellars. During winter months, cellar temperature is typically –5 to –15 °C that is much warmer than outside air temperature in Siberia (–40 to –60 °C). Cold dense air goes into the bottom of the cellar through the lead pipe. Warmer cellar air escapes through the upper lead pipe to the atmosphere. This natural cooling system would be a great advancement. We observed this type of design in Siberian indigenous communities around Baikal and Chukotka. Air convection systems work efficiently during the winter months especially after cold snaps. During late winter, the temperature gradient in ice cellars is reversed and heat removal is terminated once outside air temperature getting warmer.

As mentioned earlier, ice cellars are an important cultural and economic resource for residents of Arctic communities. Changes in climate could significantly affect the ground thermal regime around ice cellars; in addition, processes of flooding and urban impact could further be detrimental to future ice cellars. The impact of a changing environment on ice cellars will require further site-specific investigations. Soil characteristics and ground ice conditions vary substantially over distances of only a few meters, necessitating detailed surveyings. In cases where ice cellar degradation is observed, many engineering options are available for maintenance; for example, thermo-siphons could be used to artificially maintain frozen conditions (Wendler 2011). Thermo-siphons, though having a long history of use in Sakha by Russian engineers, have never been used to maintain ice cellars except an oil company trying to test in Kaktovik, Alaska.

14.5 Summary and Conclusions

Permafrost is considered as impermeable layer in the ground. However, in real world, water moves in all seasons over permafrost regions. This chapter discussed and apprehend water geometry (unfrozen or brine) and the related indigenous people's life. Since Last Glacial Maximum, earth history indicated climate warming dramatically. Especially, big hydrological events happened in early Holocene (e.g., Climatic Optimum) formed many thermokarst lakes and permafrost degradating processes. After the Little Ice Age, there is a warm event on earth and that would be continuous with more warming in the future similar to the last interglacial period. Warming permafrost implies more dynamic hydrological process and necessary adaptation of human life in the North. Open talik formation will be more expanded in continuous permafrost regions that impact aufeis formations and/or lake geometries. Simply, surface water dynamics have to consider more factors and linkages, including underground water connections. In current state, we are not certain about relationship between aufeis formations (size) and climate variability. Aufeis has quite contribution of the winter baseflow and should including total annual discharge distribution.

For Ice cellar of northern people, there is some changing culture after electric freezer invented. The purpose of using cellar is not only for food storages depending on regions and tribes, although the cellar is used for seasonal storage meats, fishes, ice (drinking water for tea in summer), or fermentation purpose. Future permafrost temperature will be slightly changing the usage and strictures of the cellars. In case of permafrost warming, it could be more seeping out cryopeg's brine water into cellar as well as hazards to flooding events. However, in general, hydrological interactions are most impacted parameter such as weakening of the structure or collapsing of the ceiling.

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