See also

Air–Sea Interaction: Momentum, Heat and Vapor Fluxes; Sea Surface Temperature. **Boundary Layers:** Complex Terrain; Convective Boundary Layer; Neutrally Stratified Boundary Layer; Stably Stratified Boundary Layer; Surface Layer. **Energy Balance Model, Surface**. **Land–Atmosphere Interactions:** Overview. **Numerical Models:** Methods.

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PERMAFROST

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Introduction

Permafrost, ground that remains at or below 0°C for two or more years, underlies about a fifth of the land surface of the Earth. Permafrost terrain consists of an active layer at the surface that freezes and thaws each year, underlain by perennially frozen ground. The top of permafrost is at the base of the active layer, and the base of permafrost occurs where the ground temperature rises above 0°C at depth (**Figure 1**).

The first scientific reports on permafrost were published in the 1830s by the Royal Geographical Society of London. These papers reported the thickness of frozen ground in a well at Yakutsk, Russia, and provided instructions to officers of the Hudson's Bay Company on describing the phenomenon. The first systematic study published in English of perennially frozen ground was prompted by strategic considerations in World War II, when the Alaska Highway was built through northern British Columbia and Yukon to Alaska, with the associated Canol Pipeline. At that time, the term 'permafrost' was coined by S. W. Müller as a contraction of 'permanent frost'.

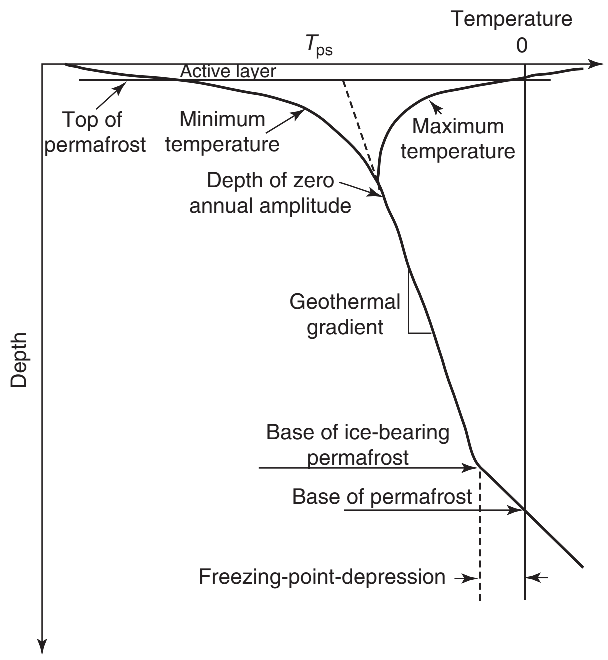

Figure 1 Schematic temperature profiles in permafrost, illustrating the annual maximum and minimum temperatures. Annual mean permafrost surface temperature, T_{ps}, is obtained by extrapolating the common linear portion of the profiles upwards. Soil particle effects, solutes, and hydrostatic pressure decrease the phase equilibrium temperature so that a layer just above the base of the permafrost does not contain ice. The change in slope of the temperature profile at the base of ice-bearing permafrost is caused by the difference in thermal conductivities between the frozen and unfrozen ground.

Permafrost grows by freezing from its base downward or, when new material is added to the ground surface, from the top upwards, or by a combination of these processes. The first permafrost on Earth likely formed prior to or during the first ice age about 2.3 billion years ago, and its occurrence, distribution, and thicknesses have varied in response to repeated ice ages throughout Earth's history.

There is scientific and geotechnical interest in permafrost principally because it contains ice, is close to its thawing point, and is sensitive to changes in surface conditions caused by human activities and climate. Temperatures in permafrost present retrievable records of past climate, and climatic change commonly leaves an imprint on the stratigraphy of ground ice. Ice-rich permafrost contains ice in excess of the water content at saturation, with the ice masses commonly distributed as lenses, millimeters to centimeters in thickness, within the ground. Massive ground ice with dimensions typically from meters to tens of meters also occurs in permafrost (**Figure 2**).

Occurrence, Distribution, and Thickness

The spatial extent of permafrost generally changes with climate, but there can be considerable regional variation because of snow cover and other factors. Over half of Canada and Russia, most of Alaska, and north-east China are underlain by continental permafrost, while alpine permafrost is found at high elevations in middle and low latitudes (e.g., the summit of Mauna Kea in Hawaii). Some permafrost is found in Scandinavia, but the spatial extent is much less than at corresponding latitudes of North America, because of the warming influence of the Gulf Stream. Permafrost containing water ice and other ices is known to exist on Mars and on other bodies in our solar system.

Permafrost regions are divided into zones with varying spatial extent of perennially frozen ground. In the continuous permafrost zone (**Figure 3**), more than 90% of the ground is underlain by permafrost, and it is usually absent only beneath rivers and lakes that do

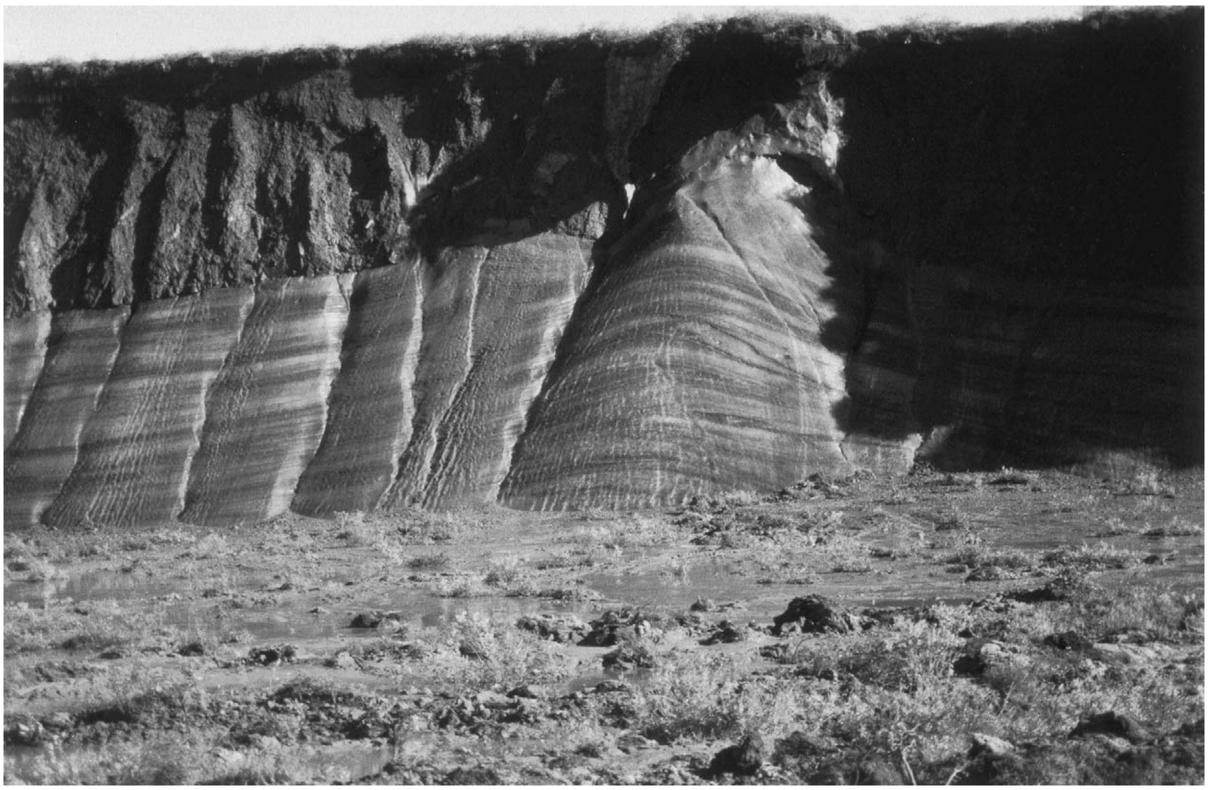

Figure 2 Massive ground ice exposed at the Beaufort Sea coast near Tuktoyaktuk, Northwest Territories. The ice was formed during permafrost aggradation after deglaciation, with water supplied from the underlying sands. Banding indicates variations in the concentration of sediment suspended in the ice. Undulation of the banding indicates displacement of the ice subsequent to formation. (Photograph by J. R. Mackay. See Mackay JR and Dallimore SR (1992) Massive ice of the Tuktoyaktuk area, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 29: 1235–1249.)

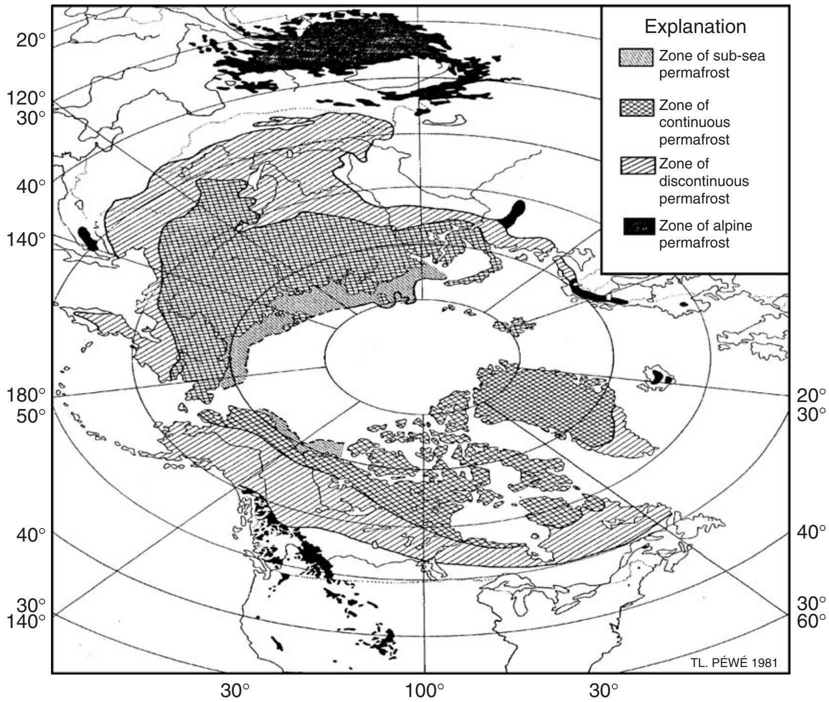

Figure 3 Generalized map of the approximate distribution of permafrost in the north circumpolar region of the Earth. Direct data (probing, drilling, sampling, temperature measurements) are scarce, so the map is somewhat unreliable, especially at local scales in discontinuous permafrost. Subsea permafrost typically exists in the continental shelves of the Arctic Ocean where seabed temperatures remain negative. (Adapted from Péwé TL (1983) Alpine permafrost in the contiguous United States: a review. *Arctic and Alpine Research* 15: 145–156, by permission of the Regents of the University of Colorado).

not freeze to their bottom in winter. At continental scale, the limiting annual mean air isotherm for continuous permafrost is about -8°C. At warmer temperatures, variations due to microclimatic effects lead to a zone of widespread discontinuous permafrost where 50 to 90% of the ground is underlain by permafrost. In the sporadic discontinuous permafrost zone, 10 to 50% of the ground is underlain by permafrost.

Subsea permafrost formed during glacial periods, when lower sea levels exposed large areas of the polar continental shelves. The cold air temperatures and long periods of exposure allowed permafrost to grow there to great depths. Thicknesses of several hundreds of meters are known to exist in these shelves and, at present, this permafrost is slowly degrading beneath the relatively warm and salty ocean. However, several tens of thousands of years are required to thaw it completely.

The distribution and thickness of permafrost are controlled by factors that influence its heat balance and heat flow within it. Since pore spaces in permafrost are generally blocked by ice, heat flow is by conduction. While the presence of permafrost is due primarily to climate, considerable modification of the temperature between the atmosphere and permafrost may occur, owing to vegetation, energy exchanges at the snow–ground surface, transfer of heat through the active layer, and local geological and hydrological conditions. In particular, the winter snow cover buffers the ground from frigid air temperatures, and the annual mean ground surface temperature, T_s, is commonly 2 to 4°C warmer than the annual mean air temperature, T_a. In the summer months, shade from vegetation and a supply of soil moisture for evaporation are two significant site variables that reduce the ground surface temperature. T_a and T_s tend to be similar at windswept sites where there is little snow accumulation. Significant changes in ground temperatures occur across the continental tree line, for within the forest there is less wind at ground level than on the tundra, and hence the snow is deeper, less dense, and a better insulator.

An organic horizon at the ground surface commonly assists the development and persistence of permafrost. Dry moss and peat have low thermal conductivities, reducing heat flow into the ground in summer. However, autumn rain characteristically increases the water content of the moss and peat, increasing their conductivity (particularly when frozen) and facilitating extraction of heat from the ground in winter. These principles indicate why, near the southerly limit for continental permafrost in North America, permafrost generally occurs in peat lands and under thick moss layers.

At regional scale, permafrost distribution may be controlled by topography, in terms both of aspect, modifying the receipt of radiation, and of cold conditions brought by high elevation. However, in the mountain and plateau regions of Alaska and adjacent Canada, strengthening of winter inversions by cold-air drainage into valley bottoms enhances conditions for permafrost there.

The thickness of permafrost is commonly influenced by geothermal heat flow and bedrock stratigraphy. At continental scale, heat flow varies by a factor of four between stable craton, where the flow is low, and tectonically active terranes. This geothermal heat is conducted through the permafrost where the temperature gradient varies with the thermal conductivity of the bedrock. At local and regional scales, the movement of groundwater carries heat that can modify the spatial distribution and thickness of permafrost. Under equilibrium conditions where a constant annual mean permafrost surface temperature, T_{ps}, has existed for a long time, the thickness of homogeneous permafrost, Z_e, is governed by T_{ps}, the thermal conductivity of permafrost, K_p, and the geothermal heat flow, J, where (from Fourier's law)

$$Z_e \approx \frac{K_p}{J} T_{ps} \qquad [1]$$

Equation [1] provides reasonable estimates of Z_e when T_{ps} remains near its long-term value or has been close to it for a sufficient period. While T_a is the principal variable governing T_{ps}, the values differ owing to the effects of snow cover, the active layer, and other factors mentioned above.

Properties and Processes

Unfrozen Water and Ice

Seasonally frozen ground and permafrost contain unfrozen water (**Figure 4**) and ice in equilibrium at temperatures less than 0°C as a result of the effects of soil particles and solutes. In the absence of solutes, temperature, T, and the soil's specific surface area, S, are the primary determinants of the amount of unfrozen water, θ_u, and are empirically related by

$$\ln \theta_u = 0.2618 + 0.5519 \ln S - 1.449 S^{-0.264} \ln |T| \qquad [2]$$

where θ_u is in percent. Unfrozen water reduces ground thermal conductivity and distributes latent heat over a range of temperatures, so that temperature changes require freezing or thawing throughout the frozen ground. These effects retard the thermal response of the active layer and permafrost.

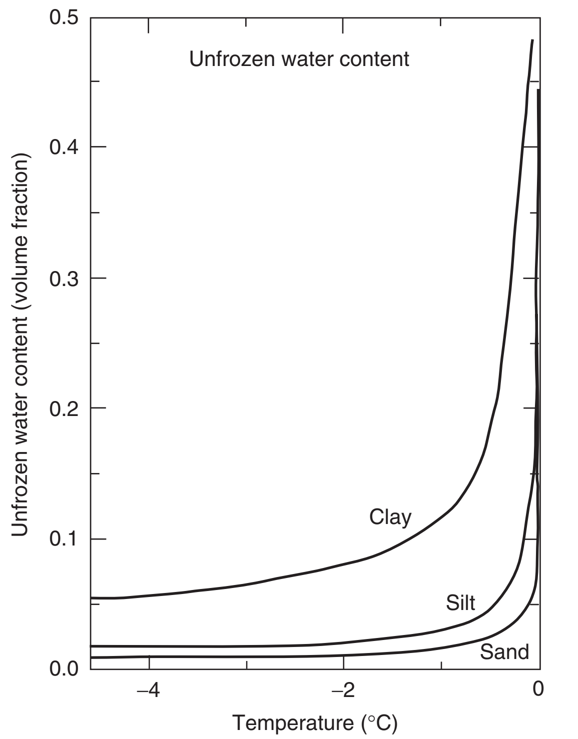

Figure 4 Representative values for the temperature dependence of unfrozen water contents in sand, silt, and clay. Unfrozen water contents typically increase with temperature and finer-grained soil and are small in moss and peat. The presence of solutes increases unfrozen water contents.

Frost Heave

Unfrozen water is also responsible for the development of ice lenses in the ground during freezing, leading to local uplift (heave) of the surface. The water occurs in mobile films on particle surfaces, where it is held in tension. Water flows along the tension and thermal gradients in the films to cooler regions of the ground. When the water content in the freezing soil exceeds saturation, the excess volume separates soil particles to form layers, or lenses, of segregated ice. Ice segregation and frost heave characteristically occur in fine-grained soils, where the unfrozen water content is sufficiently large to conduct water into the freezing ground. In coarse-grained soil, the unfrozen water content is small and the permeability is too low to allow water migration during freezing. In saturated, coarse-grained soil, the expansion of water during freezing is accommodated by expelling the excess water into unfrozen ground ahead of the freezing front. When the freezing system is closed, pore-water expulsion may also lead to heave as hydrostatic pressure deforms the overlying frozen ground.

Active Layer

Permafrost is separated from the atmosphere by a boundary layer consisting of the active layer and vegetation in summer with the snow cover added in winter. The active layer transmits heat to and from permafrost, reduces the amplitude of thermal variations at the top of permafrost compared with the ground surface, is the medium through which moisture and gases are exchanged between the permafrost and the atmosphere, and provides water and nutrients for biological processes. In permafrost terrain, the active layer supports plant and animal communities since virtually all biological activity below ground occurs within it.

Permafrost immediately below the active layer is characteristically ice-rich. This ice-rich zone acts as an impermeable barrier to drainage, so much permafrost terrain is wet. The ice-rich zone is the reason permafrost terrain is considered sensitive to disturbance, for deepening of the active layer commonly leads to subsidence (thaw settlement) and accelerated erosion as the ice melts.

In a dry active layer, $T_s \approx T_{ps}$; however, in a wet one, there is an asymmetry in heat flow because the frozen thermal conductivity, K_f, exceeds K_t, the thawed conductivity. This makes T_s warmer than T_{ps} and the difference is the thermal offset (**Figure 5**)

$$T_{ps} - T_s = \frac{I_t}{P}\left(\frac{K_t}{K_f} - 1\right) \qquad [3]$$

where I_t is the thawing index at the ground surface and P is the period (365 days).

Once the snow melts, the ground surface warms above 0°C and thawing of the active layer begins (**Figure 6**). In a simplified model, its maximum thickness is

$$X \approx \sqrt{\frac{2K_t I_t}{h}} \qquad [4]$$

where h is the volumetric latent heat of the ground, which depends on the ice content. The thickest active layers (1 to 2 m or more) develop in dry materials, characteristically bedrock, sand, and gravel. Thin active layers are common in wet organic soils, where there may be considerable amounts of ice at the beginning of summer, and where the thermal conductivity of a dry surface layer is very low. The thinnest active layers, less than 30 cm thick, occur in the High Arctic, owing to short, cool summers.

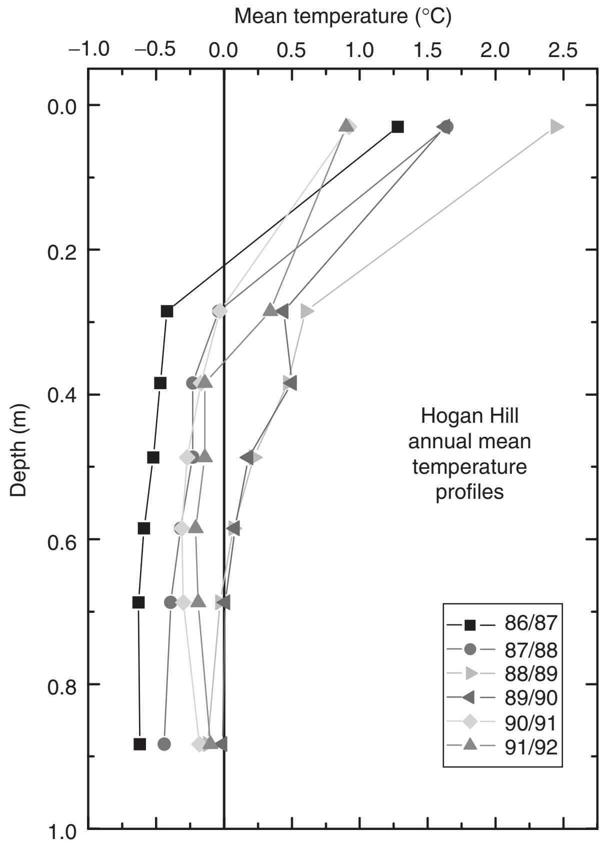

Figure 5 Annual mean temperature profiles in the active layer showing the effects of thermal offset. In the context of climatic change, thermal offset allows permafrost to survive changes that produce annual mean ground surface temperatures warmer than 0°C over multiyear periods as shown at this site. (Adapted with permission from Osterkamp TE and Romanovsky VE (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes* 10: 17–37.)

The active layer typically reaches X and begins freezing upward from the bottom when $T_s > 0$°C, one or two weeks before freezing downward from the surface. Temperature changes during freezing are retarded by the presence of unfrozen water for a few weeks in cold permafrost and for much of the winter in warm permafrost. The rate of freezing down from the surface depends on the amount and timing of the early winter snow cover and on the thermal conductivity and moisture content of the active layer. Freezing may be slow if a deep snow cover is established early in winter on a wet active layer. Temperatures in the portion of the active layer that remain thawed are constrained near 0°C, a phenomenon termed the 'zero curtain', until freezing is complete.

Thermal Regime

A temperature profile in permafrost that is homogeneous, does not contain unfrozen water, and is in

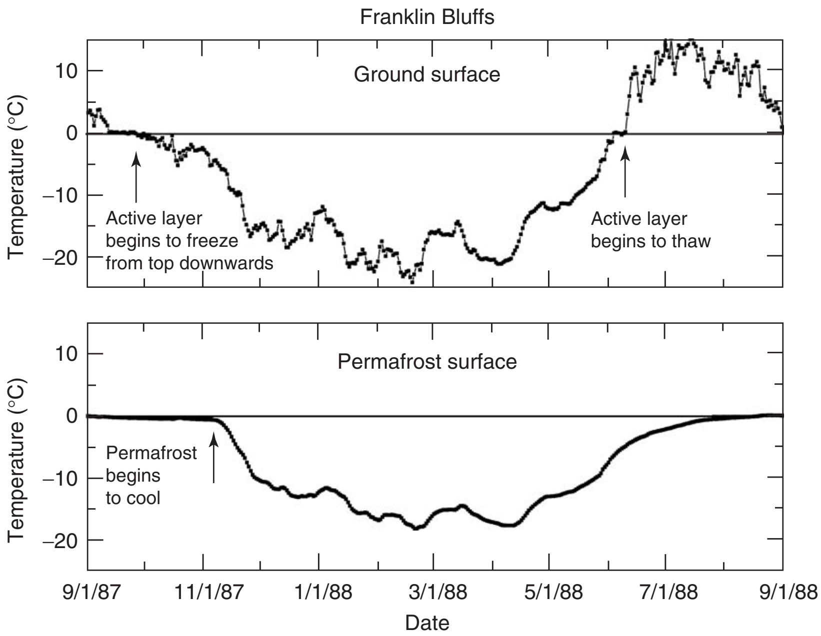

Figure 6 Time series of temperatures at the ground and permafrost surfaces for the annual cycle. Freezing of the active layer from the top downward began about 26 September 1987, and from the bottom upward about ten days earlier. The zero curtain persisted until about 12 November 1987, the date of freeze-up of the active layer. At this time, the lower portion of the active layer and upper portion of the permafrost began to cool. Ground surface temperatures during spring, 1988, remained near 0°C from 4 to 11 June (the period of snowmelt), when the active layer began to thaw.

equilibrium with a constant long-term T_{ps}, is a straight line described by Fourier's law (seasonal variations near the surface and freezing point depression near the base cause deviations from a straight line). However, the thermal regime of permafrost is often characteristically different during its formation and growth, after surface temperatures change, and during thawing. Prior to permafrost formation, interannual variability in T_s and in conditions within the seasonally frozen layer can cause the depth of freezing to exceed the depth of thawing during the following summer, resulting in a temporary layer of frozen ground. T_s can be significantly positive when this occurs, owing to thermal offset. A long-term shift in conditions allows such a layer to persist and grow, creating a thickening layer of permafrost with time. When the permafrost freezes from its base downward, the solution of the Stefan problem yields the approximate depth to the bottom of the growing permafrost at time t:

$$Z \approx \sqrt{\frac{2K_p(-T_{ps})t}{h}} \quad [5]$$

In eqn [5], K_p and h depend on ground properties. Equation [5] neglects geothermal heat flow and the effects of freezing point depression near the base of permafrost (**Figure 1**). It reduces to $Z \approx \text{constant} * \sqrt{t}$, where the constant is typically about 1 to 5 for a wide range of conditions, where t is measured in years and Z in meters.

In principle, permafrost grows until $Z = Z_e$, when it is in equilibrium with T_{ps}. However, in reality, T_{ps} is variable over the long time scales needed to grow even relatively thin permafrost, resulting in permafrost thicknesses that vary with long-term variations in T_{ps} (**Figure 7**).

A change in T_{ps} (warming or cooling) produces a curvature in the temperature profile that penetrates deeper with time (**Figure 8**). The magnitude of the change at any depth can be obtained from the measurements and the timing can be calculated from the maximum depth of penetration of the thermal signal. For permafrost with thickness Z, the time scale required for the temperature profile to respond to a new surface condition is

$$\tau \approx \frac{Z^2}{4D} \quad [6]$$

where D is the thermal diffusivity of the permafrost. Freezing or thawing at the bottom of the permafrost begins once the thermal disturbance has penetrated there, with calculated rates in thick continuous permafrost that are typically millimeters per year (**Figure 7**).

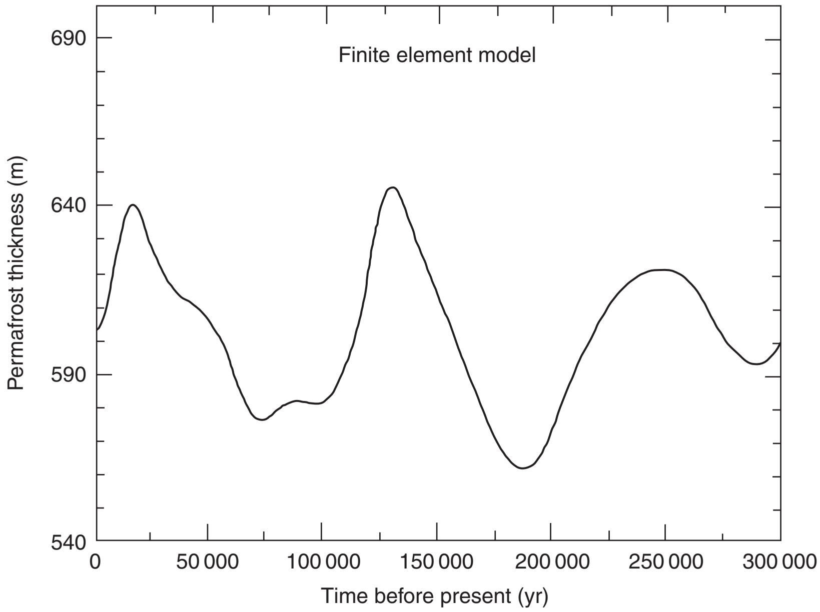

Figure 7 Calculated permafrost thickness variations in response to changes in paleoclimate. Thicknesses varied about 84 m (562 m $\leq Z \leq$ 646 m) with maximum thawing rates of 5 mm yr^{-1} and freezing rates of 2 to 3 mm yr^{-1}. (Adapted with permission from Osterkamp TE and Gosink JP (1991) Variations in permafrost thickness in response to changes in paleoclimate. *Journal of Geophysical Research* 96: 4423–4434. Copyright © by the American Geophysical Union.)

When $T_{ps} \approx 0°C$, interannual variability in T_s and conditions within the active layer may cause the depth of summer thaw to exceed the depth of winter freezing, resulting in a temporary residual thaw layer between seasonally frozen ground and permafrost. T_s can be significantly positive when this occurs, and interannual variability may allow the layer to refreeze and thaw repeatedly. If the conditions that caused the warming persist, the layer may become too thick to refreeze. Then, the permafrost is decoupled from the atmosphere and warms continuously throughout the year. An equation similar to [5] can be obtained for the thickness of the thawed layer above permafrost. Thawing rates at the top of the permafrost depend on the temperatures at the bottom of the former active layer, K_t, and h and can be a few tenths of a meter per year for near-surface permafrost. While the permafrost is thawing, temperatures within it warm very slowly as they approach 0°C (**Figure 9**) because of the effects of unfrozen water.

Geomorphic Features

There are three principal geomorphic features unique to permafrost terrain: pingos (conical ice-cored hills) and ice-wedge polygons, both associated with permafrost aggradation; and thermokarst terrain, associated with ground thawing. Pingos may form in the unfrozen sandy sediments of drained-lake bottoms that are completely surrounded by permafrost. After drainage, lake sediments freeze primarily from the top downward and from the sides and bottom of the talik (the unfrozen layer) inward. Freezing of the sands results in pore-water expulsion into the enclosed talik, increasing hydrostatic pressure there. The increased pressure lifts the permafrost, and the water freezes in place, to create a core of ice in the mound (**Figure 10**). Pingos grow until the talik freezes completely. There are 1350 pingos in Canada's western Arctic, and the largest, Ibyuk Pingo, is 50 m high, over 1200 years old, and is growing about 2 cm taller annually. Pingos are also found at the base of hill slopes, where the pressure to lift permafrost may be provided by confined groundwater flowing down slope.

Ice-wedge polygons are ubiquitous in continuous permafrost terrain, where cracks occur in the permafrost as a result of thermal contraction in winter. The cracks relieve thermal stress approximately normal to their axes, yielding a polygonal network that may subdivide further as more cracks form. Ice wedges begin to grow in the cracks when spring snowmelt infiltrates them and freezes to form near-vertical sheets of ice. Repetition of the process over many years leads to development of wedges of ice, which commonly extend downwards from the base of the active layer for about 4 m (**Figure 11**), although some exceed 10 m in depth. Polygons are clearly expressed in lowlands, but movement of the active layer commonly obscures

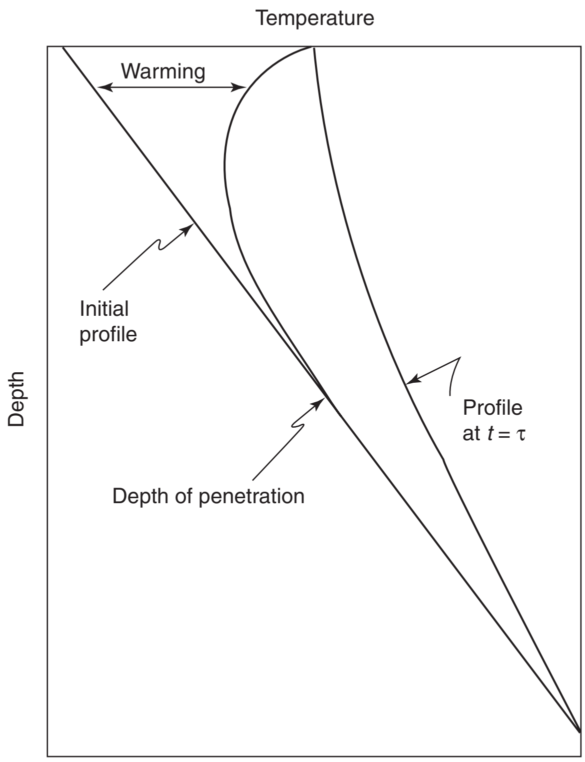

Figure 8 Schematic thermal response of permafrost to warming of its surface temperature. Seasonal variations near the surface are not shown. The magnitude of the warming at any depth can be obtained directly from the measurements and the timing of the warming can be calculated from the depth of penetration of the warming signal. (Adapted with permission from Lachenbruch AH, Sass JH, Marshall BV, and Moses TH (1982) Permafrost, heat flow, and the geothermal regime at Prudhoe Bay, Alaska. *Journal of Geophysical Research* 87: 9301–9316. Copyright © by the American Geophysical Union.)

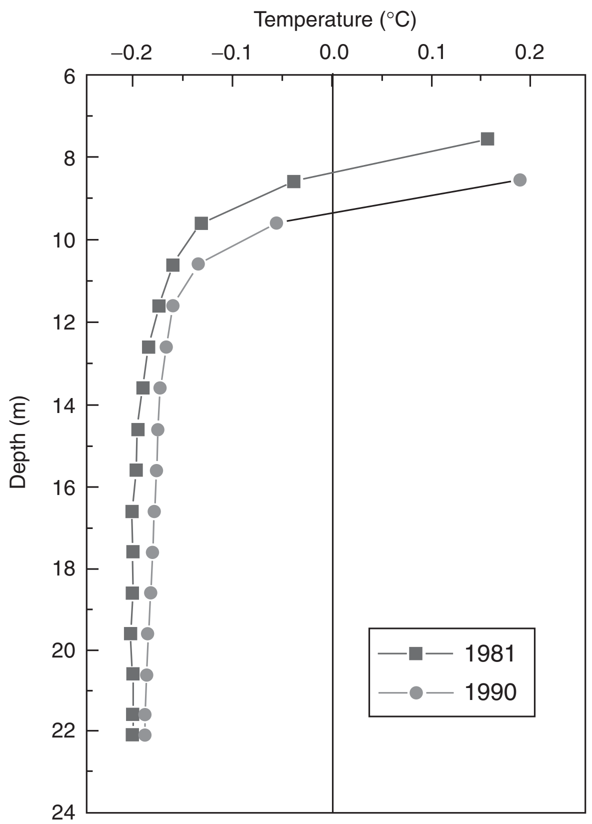

Figure 9 Temperature profiles in thawing permafrost near Fairbanks, Alaska obtained 9 years apart. The presence of unfrozen water retards temperature changes and produces the observed curvature below 8 m depth. (Adapted with permission from Osterkamp TE and Romanovsky VE (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes* 10: 17–37.)

them on hill slopes. Ice-wedge growth deforms the surrounding ground to accommodate the additional volume. Growth of the ice wedges often forces the adjacent ground upwards and laterally, creating a trough above the ice wedge.

Warming or disturbance to permafrost terrain usually leads to deepening of the active layer and thawing of the ice-rich zone at the top of permafrost or of near-surface massive ground ice, causing local subsidence. This thaw settlement produces a pitted relief, called thermokarst terrain. Drainage conditions determine whether standing water will be present or not. When depressions in thermokarst terrain collect water, the disturbance to permafrost is enhanced, leading to growth of thermokarst lakes. ‘Beaded’ streams occur when a series of pools (beads) form along the stream as a result of thawing ice-rich permafrost or large ice masses. Most features of thermokarst terrain range from 1 to 100 m in lateral dimensions, although thermokarst lakes are often larger. Thermokarst depressions up to 100 km^2 in area and 5 to 20 m in depth form part of the landscape of Siberia, where they are called ‘alasses’.

Thermokarst terrain also includes retrogressive thaw slumps (**Figure 12**) that commonly develop where ice-rich permafrost is exposed by erosion in riverbanks, lakeshores, and along the coast, or by other processes. These features, with a near-vertical retreating headwall and a low-angled foot slope, are the commonest form of landslide in permafrost terrain. Landslides involving only the thawing active layer also occur, particularly in areas of fine-grained, ice-rich soil, when the active layer and vegetation detach from the underlying frozen material. If ice-rich permafrost or massive ground ice is exposed, the slope failure may develop into a thaw slump.

Impacts of Climatic Warming

The Earth’s climate has generally warmed since the mid-to late 1800s. In the Arctic and Sub-Arctic,

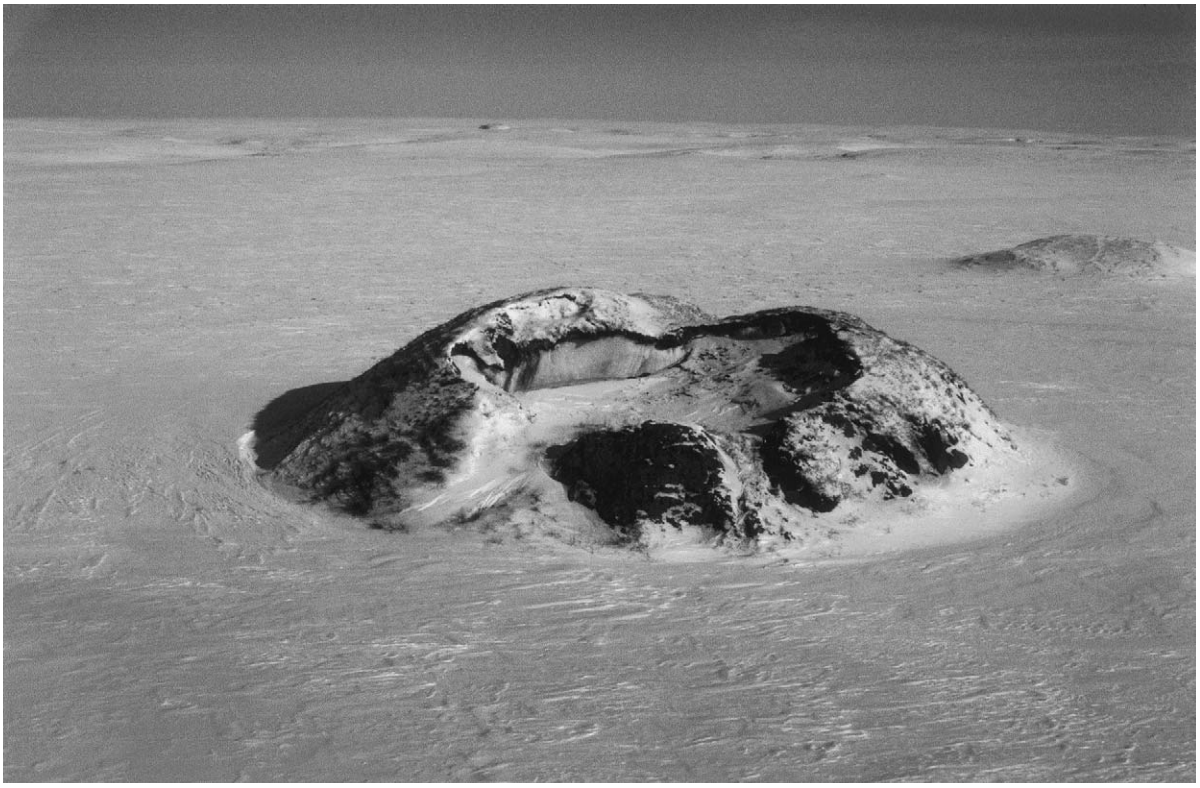

Figure 10 Pingo near Wolf Lake, Richards Island, western Arctic coast, Canada, about 20 m high, with the ice core exposed after collapse of the central portion. The vertical extent of exposed ice in the far headwall is about 4 m. (Photograph by C. R. Burn. See Mackay JR (1998) Pingo growth and collapse, Tuktoyaktuk Peninsula area, western Arctic coast, Canada: a long-term field study. *Géographie physique et Quaternaire* 52: 271–323.)

permafrost has also warmed, particularly in Russia, China, Mongolia, Alaska, and western Canada. In Alaska, T_{ps} for continuous permafrost warmed 2 to 4°C over the last century followed by a cooling in the early 1980s and then a warming of up to 3°C since then. In central Alaska and the Yukon Territory of Canada, discontinuous permafrost has warmed since 1970 as a result of changes in air temperatures and snow cover. While it has warmed typically 1 to 2°C since the late 1980s in Alaska, the ground has been cooling in the Yukon owing to a reduction in snow cover. In the Northwest Territories, warming of permafrost by about 1°C has been observed since the early 1970s, but in northern Québec there has been cooling. Such regional variations are to be expected as a result of spatial climatic variability. However, the general signal from the Northern Hemisphere indicates permafrost warming, and in some areas permafrost is presently thawing at both top and bottom, the southern boundary of the permafrost is moving northward, and the occurrence of thermokarst terrain is increasing.

Current global circulation models predict that air temperatures in the Arctic will rise 2 to 5°C during the next half-century, with the warming in winter larger than in summer and with more precipitation throughout the year. While there is considerable uncertainty in these predictions, warming of T_{ps} by a few degrees would have serious consequences for permafrost regions, especially for discontinuous and sporadic discontinuous permafrost.

In the continuous zone, the active layer, thaw lakes, coastal processes, landscape processes, eolian activity, and vegetation would be sensitive to climatic warming. The effect on the permafrost would be to warm it and, possibly, to change the depth of the active layer. Thawing at the base of the permafrost would start several centuries or more later.

Since most of the discontinuous and sporadic discontinuous permafrost is within a few degrees of thawing, it can be expected to warm, to begin thawing from the top and bottom with an increase in the incidence of thermokarst terrain, and, eventually, to disappear. Although many centuries would be required for the permafrost to disappear, thawing of the warmest permafrost from the top would begin immediately.

Thawing of permafrost as a consequence of human activities and as a result of the climatic warming since the late 1800s serves as a model for what may be

Figure 11 Ice wedge exposed in a coastal bluff near Tuktoyaktuk, Northwest Territories, Canada, about 4 m tall. The photograph was taken in May 1992, when the winter's thermal contraction crack was visible. (Photograph by C. R. Burn. See Mackay JR (2000) Thermally induced movements in ice-wedge polygons, western Arctic coast: a long-term study. *Géographie physique et Quaternaire* 54: 41–68.)

expected to occur with additional climatic warming. Where permafrost contains massive ground ice or is ice-rich, extensive differential thaw settlement has occurred, with damage to the natural terrain and to infrastructure. Human-induced thaw settlement is at present responsible for damage to infrastructure on permafrost where the built structures have raised ground temperatures above 0°C (**Figure 13**). The magnitude of the thaw settlement is typically 1 to 3 m, but can exceed 5 m (vertical settling) of the ground surface. Repair of the infrastructure is costly and some structures, airports, and roads have been abandoned.

Carbon and Trace Gases

Soils in northern regions may contain one-quarter to one-third of the Earth's total soil carbon pool, with much of it stored in frozen peat in near-surface permafrost. When permafrost thaws, the carbon is cycled through terrestrial ecosystems with the gaseous end products (CO_2 and CH_4) emitted to

Figure 12 Headwall of a retrogressive thaw slump in ice-rich glaciolacustrine sediments near Mayo, Yukon Territory, Canada. Lenses of segregated ice give the exposure its texture. The ice lenses developed when permafrost formed in glacial lake sediments after drainage of the lake at the end of the last glaciation. A deep active layer, which formed during the warmest climate of the last 10 000 years, left a thaw unconformity in the sediments marked by the abrupt transition between the darker, lower, ice-rich material, and the lighter sediment above. (Photograph by C. R. Burn. See Burn CR (2000) The thermal regime of a retrogressive thaw slump near Mayo, Yukon Territory. *Canadian Journal of Earth Sciences* 37: 967–981.)

the atmosphere. While the details of this process are not well understood and are subject to environmental constraints, there is evidence that tundra regions may have shifted from being a carbon sink to a carbon source.

Gas Hydrates

Temperature and pressure conditions within and under thick permafrost are favorable for the formation and existence of gas hydrates that are a potentially abundant source of energy. Warming of the permafrost would eventually destabilize these hydrates, producing large quantities of gas (primarily CH_4) that may find its way into the atmosphere. Long time scales (many centuries or millennia) are required for this process. However, subsea permafrost in the continental shelves of the Arctic Ocean that was submerged and warmed by seawater more than a few thousand years ago could be emitting methane at present. Currently, there is too little information to adequately assess this problem.

Ecosystems

In areas of ice-rich permafrost, thaw settlement and development of thermokarst terrain destroys the substrate on which the current ecosystems rest, dramatically changing the nature of the ecosystems. The effects have been observed to include:

1. Destruction of trees and reduction in area of boreal forest ecosystems.
2. Expansion of thaw lakes, grasslands, and wetlands.
3. Destruction of habitat for caribou and terrestrial birds and mammals.
4. Formation of new habitat for aquatic birds and mammals.
5. Coastal and riverbank erosion.
6. Clogging of salmon spawning streams with sediment and debris.
7. Slope instabilities, thaw slumps, landslides, and erosion.
8. Talik development, with increasing depth to water table.
9. Increased methane emissions in wet areas.

Figure 13 Longitudinal cracks in a road embankment near Fairbanks, Alaska. Snow removal during winter produces a berm along the shoulder and slope of the embankment that warms the underlying ice-rich permafrost, causing it to thaw and settle. Settlement under the shoulder and slope causes the edge of the embankment to tilt outward, putting the top surface of the embankment in tension, which eventually results in cracks. Patches (darker asphalt) in the pavement are a result of two long cracks and the guardrail on the far side of the road has sagged as a result of the thaw settlement. (Photograph by T. E. Osterkamp.)

Thermokarst has been observed to result in the partial or complete destruction of some ecosystems and their conversion to other types of ecosystems. In one lowland area in central Alaska, permafrost degradation is widespread and rapid, causing large shifts in ecosystems from birch forests to fens and bogs (**Figure 14**). If current conditions persist, the remaining birch forests will be eliminated by the end of the twenty-first century.

Nomenclature

- D thermal diffusivity ($m^2\,s^{-1}$)
- h Volumetric latent heat of the sediments ($J\,m^{-3}$).
- I_t Thawing index at the ground surface (usually stated in °C days).
- J Geothermal heat flow, negative for a positive temperature gradient, ($W\,m^{-2}$).
- K_p Thermal conductivity of permafrost, ($W\,m^{-1}\,K^{-1}$).
- K_f Thermal conductivity of the frozen active layer, ($W\,m^{-1}\,K^{-1}$).
- K_t Thermal conductivity of the thawed active layer, ($W\,m^{-1}\,K^{-1}$).
- P Period of the annual temperature wave, (365 days).
- S Soil specific surface area, ($m^2\,g^{-1}$).
- t Time, (s).
- T Ground temperature, (°C).
- T_s Annual mean temperature at the ground surface, (°C).
- T_{ps} Annual mean permafrost surface temperature, (°C).
- X Maximum thickness of the active layer, (m).
- Z Thickness of permafrost at time, t, (m).
- Z_e Equilibrium thickness of permafrost, (m).

Figure 14 An area in the Tanana River valley near Fairbanks, Alaska, showing ponds, floating fens and remnant birch forest underlain by ice-rich permafrost. Thawing is resulting in complete destruction of the trees and forest ecosystem, which is being converted into floating fens with ponds. Standing dead birch trees, some on ground that has settled below the water level, are visible in the center and left half of the picture. (Photograph by M. T. Jorgenson. Adapted with permission from Osterkamp TE, Viereck L, Shur Y, *et al.* (2000). Observations of thermokarst in boreal forests in Alaska. *Arctic, Antarctic, and Alpine Research* 32: 303–315 by permission of the Regents of the University of Colorado).

θ_u Unfrozen water content, (kg of water per kg of dry soil, sometimes stated in percent).

τ Time scale for a permafrost temperature profile to respond to a new surface temperature, (usually stated in years).

See also

Arctic Climate. **Biogeochemical Cycles:** Carbon Cycle. **Cold Air Damming**. **Energy Balance Model, Surface**. **Global Change:** Biospheric Impacts and Feedbacks; Human Impact of Climate Change; Surface Temperature Trends. **Hydrology:** Soil Moisture. **Ice Ages (Milankovitch Theory)**. **Methane**. **Reflectance and Albedo, Surface**. **Satellite Remote Sensing:** Aerosol Measurements; Precipitation; Temperature Soundings. **Snow (Surface)**.

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