State of the Earth's Cryosphere at the Beginning of the 21st Century: Glaciers, Global Snow Cover, Floating Ice, and Permafrost and Periglacial Environments—

PERMAFROST AND PERIGLACIAL ENVIRONMENTS

By J. ALAN HEGINBOTTOM, JERRY BROWN, OLE HUMLUM, *and* HARALD SVENSSON

SATELLITE IMAGE ATLAS OF GLACIERS OF THE WORLD

Edited by RICHARD S. WILLIAMS, JR., *and* JANE G. FERRIGNO

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Permafrost or perennially frozen ground is defined as ground that remains at or below 0°C for at least two consecutive years. In the Northern Hemisphere, the permafrost regions occupy approximately 23×10^6 km^2 (24 percent), with actual permafrost areas underlying approximately 13 to 18 percent of the exposed land area. In the Southern Hemisphere, permafrost occurs in the Andean mountains of South America and beneath ice-free land south of 60°S. latitude. Periglacial environments are cold, non-glacial environments without tree cover and with significant frost action. A key characteristic of permafrost is the presence of ice in the ground. The ice occurs in a variety of forms and often in large masses of relatively pure ice. Numerous studies confirm widespread warming of permafrost and associated thawing of the uppermost, ice-rich permafrost. Slope failures, thermokarst, and coastal erosion are common permafrost processes. Permafrost and northern peatlands contain large quantities of carbon. Projections estimate that many tens of billions of metric tons of methane may be released as permafrost thaws. Several ground-based, global observational systems are monitoring changes in the active layer and permafrost temperatures.

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PERMAFROST AND PERIGLACIAL ENVIRONMENTS

By J. A. HEGINBOTTOM,[1] J. BROWN,[2] O. HUMLUM,[3] *and* H. SVENSSON[4]

Abstract

Permafrost is defined as ground that remains at or below 0°C for at least two consecutive years. In the Northern Hemisphere, regions of such perennially frozen ground occupy approximately 23×10^6 km^2 (24 percent of the exposed land area), with actual permafrost areas underlying an estimated 12×10^6 to 17×10^6 km^2, or approximately 13 to 18 percent of the exposed land area, principally at high latitudes. Areas beneath ice sheets and ice caps are not included in this estimate. In the Southern Hemisphere, permafrost underlies some 100 to 150×10^3 km^2 in the Andes of South America as well as lying beneath an estimated 49×10^3 km^2 of ice-free land south of lat 60°S. A key characteristic of permafrost is the presence of ice in the ground. The ice occurs in a variety of forms and often as relatively pure ice in large masses. Typically near its melting point, the widespread presence of ice in permafrost dominates the character and properties of frozen ground and dictates every engineering use of the terrain in the permafrost region. Periglacial environments are cold and nonglacial, without tree cover and with substantial frost action. The extent of permafrost corresponds broadly with the distribution of periglacial environments, although the periglacial realm is more extensive than the permafrost regions and is dominated by intense freezing and thawing of the ground. The extent and distribution of the permafrost and of periglacial regions are anticipated to decrease in response to climate warming, with consequences for natural ecosystems and for social and economic activities. Several global observational systems, including ground-based systems, remote-sensing systems, and geographic information systems (GIS) are monitoring the changes.

Introduction

Permafrost constitutes one of the major components of the perennial cryosphere, complementing ice sheets, ice caps, and other glaciers on land; snow patches; and the semipermanent pack ice and ice shelves of the polar oceans. Ephemeral components of the cryosphere include seasonal snow cover on land (and on ice sheets, ice caps, and other glaciers), seasonal ice cover on rivers, lakes, and seas, and seasonally frozen soils within and beyond the permafrost regions. Unlike the other cryospheric components, permafrost is defined strictly on the basis of ground temperature. Although definitions and terminology have varied historically (Péwé, 1974), the commonly accepted defini-

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tion today (2012) is that adopted by the International Permafrost Association (IPA): *"Ground (soil or rock and included ice or organic material) that remains at or below 0°C for at least two consecutive years"* (van Everdingen, 1998). Throughout this report, the terminology we use follows this recommendation of the IPA which, in turn, was derived from several national glossaries (van Everdingen, 1998 [revised 2005]).

Between the permafrost and the ground surface is the *active layer* (fig. 1), typically about 20 to 150 cm in thickness, which thaws during the summer and refreezes each winter (Shiklomanov and Nelson, 2007). Should the winter freezing temperature (at or below 0°C) not penetrate to the underlying permafrost, the residual unfrozen layer is called a *talik*. Taliks also occur beneath bodies of water such as deep lakes (fig. 2) and rivers. Seasonal freezing and thawing of ground without permafrost occur within the permafrost regions as well as in extensive areas beyond the permafrost

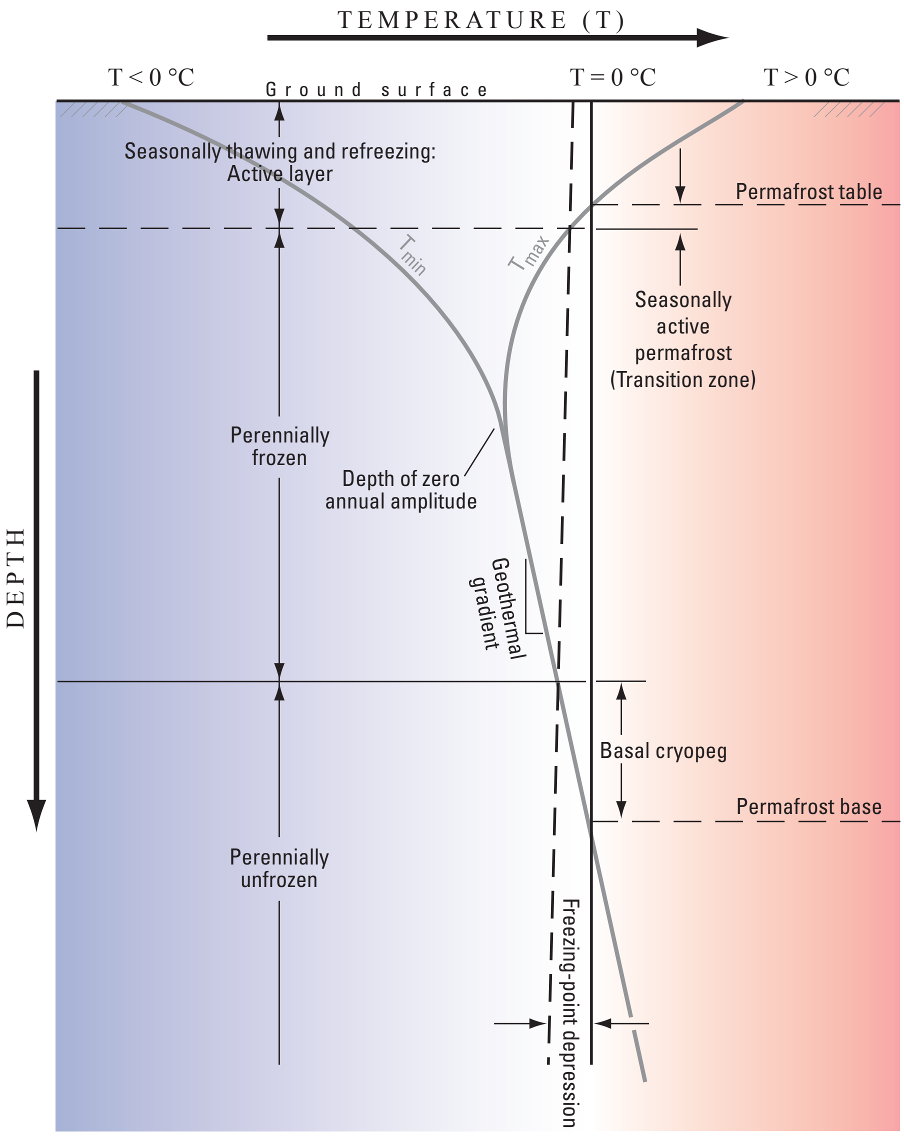


Figure 1.—*Terms used to describe ground temperature relative to 0°C in a permafrost environment (modified from van Everdingen, 1985).*

regions. The long-term average maximum extent of seasonally frozen ground in the Northern Hemisphere is estimated to be 48 million km^2 (Zhang and others, 2003).

The term *periglacial* (Lozinski, 1912) is regarded as synonymous with cold but nonglacierized surface environments that are dominated by frost-related processes or characterized by permafrost or both (French, 2003). Although many periglacial environments are underlain by permafrost, its presence is not a prerequisite. Natural tree lines often delimit the extents of periglacial environments, the tree lines representing very important ecological boundaries. In high-relief areas this boundary is well defined, more or less coinciding with a mean air temperature of 10°C for the warmest month. In low-relief regions, however, the 'tree line' may take the form of a zone 50 to 100 km in width (French, 1996), known in Russia as "the tundra–taiga boundary" and in North America as "the tundra–boreal forest transition zone" (Callaghan and Kling, 2002).

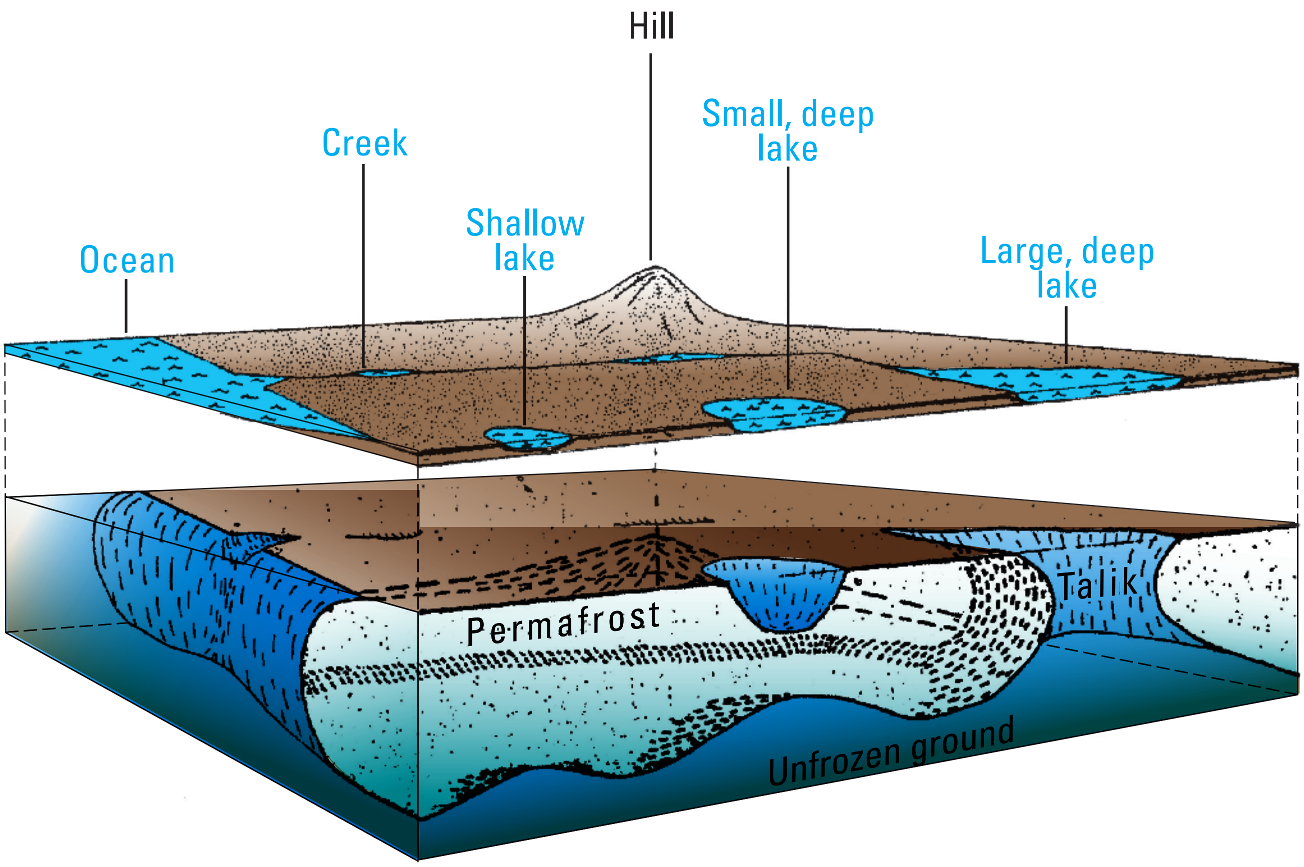


Figure 2.—*The thermal influence of water bodies on the underlying permafrost. The talik, or unfrozen layer, develops under a deep lake (modified from Lachenbruch and others, 1962)*

Although glaciers, snow cover, and floating-ice elements of landscapes are easily discernible to the human eye and on satellite imagery, permafrost terrains are not as easily detected from their surface characteristics. Indirect indicators of contemporary and paleo-permafrost include sorted and nonsorted patterned ground, pingos, ice-wedge polygons, and thermokarst features; such surface features range in size from small pitted terrain to large deep lakes. Ice-wedge polygons (Black, 1976), pingos (Müller, 1959; Flemal, 1976), and rock glaciers (Humlum, 1998) are indicators of an underlying, perennially frozen substrate. Some periglacial features are observable on high-resolution imagery and thus can be used as indicators of permafrost-dominated terrains.

This report does not attempt to present a comprehensive review of the world's literature on permafrost and related periglacial environments. The authors cite only selected references, principally those that best demonstrate the characteristics of permafrost and perigacial features as components and indicators of the cryosphere. The authors have given preference to the more easily accessible English-language publications, including the proceedings of the nine international conferences on permafrost and the journal "Permafrost and Periglacial Processes." Because the scientific literature on permafrost and periglacial environments in the Russian and Chinese languages is extensive, these English-language publications, which include many relevant and valued Russian and Chinese contributions, are of great importance. The Russian journal "Earth Cryosphere" and the Chinese "Journal of Glaciology and Geocryology" regularly report on current permafrost investigations.

Permafrost and Periglacial Environments

Brief Historical Background

The scientific study of permafrost ("ever-frozen ground" as it was then called) began in Russia in the mid-19th century (Ermolaev, 1932; Yershov, 1998; Tsytovich, 1966; Shiklomanov, 2005). In North America, encounters with permafrost began in the late 19th century, in part associated with the mining of gold (Legget, 1966). Attempts to map the areal extent of the permafrost region date from 1882 in Russia and from 1913 in North America (Nikiforov, 1928; Baranov, 1959). The existence of high-altitude permafrost in high-mountain ranges and the explanation for the existence of permafrost were recognized in Russia as early as 1757 by M.V. Lomonsov (Yershov, 1998, p. 18). However, permafrost was regarded primarily as a high-latitude phenomenon. In the early 1900s, Leffingwell (1919) investigated and reported in great detail on the origin of underground ice in the coastal regions of northern Alaska, and he provided a detailed account of the early and extensive Russian literature. Sumgin's "Permafrost Within the USSR" (Sumgin, 1927) marked the beginning of permafrost (geocryology) as a separate branch of the geological sciences (Tsytovich, 1966). In North America, military construction and engineering activities during World War II led to the publication of "Permafrost or Permanently Frozen Ground and Related Engineering Problems" by Siemon Muller (Muller, 1947, 2008; French and Nelson, 2008). The history of permafrost investigations in China is reviewed by Zhang (2005). During the latter half of the 20th century, exploration and development of petroleum and mineral resources in the Arctic, their transportation to southern markets, and military activities in the Arctic and the Subarctic led to new advances in constructing buildings on permafrost terrain (Brown, 1967).

The term *periglacial* can be traced to Walery Lozinski (Lozinski, 1912) and to participants in the excursion to Spitsbergen in 1910–11, during the XI International Geological Congress, in order to designate precisely the typical climatic and morphological environments for regions peripheral to the Pleistocene Epoch ice sheets, especially the environment in the mountainous areas of central Europe. The term thus coined was used to characterize relict features that indicated formerly active, cold-climate, geomorphological processes. The term *periglacial* was later used in a broader sense to include all varieties of frost action, especially freeze-thaw processes in cold-climate regions, except those directly connected with glaciers. Research related to periglacial geomorphology expanded exponentially during the two decades following World War II. By the mid-1960s, periglacial geomorphology was recognized as a descriptive branch of climatic geomorphology (Tricart, 1963; Dylik, 1964; Péwé, 1969) having two broad subcategories: Pleistocene Epoch and Quaternary Period studies, dealing with the mid-latitudes, and current process studies, conducted in the Subarctic and Arctic regions of North America, Scandinavia, the mountainous regions of the Southern Hemisphere, and Antarctica. Quantitative, process-oriented studies aimed at understanding periglacial landforms did not begin for several more decades (French, 2003). Today (2012), periglacial geomorphology is seen as a subdiscipline of geomorphology ("physical geography" in Europe), concerned with the landforms and processes of the cold, nonglacial regions of the world (French and Thorn, 2006).

In 1963, during the era of the Cold War, the first in a continuing series of international permafrost conferences was held at Purdue University, West Lafayette, Indiana (Woods and Leonards, 1964), bringing together engineers and scientists from all countries having an interest in permafrost and periglacial processes (Brown and Walker, 2007). In 1983, during the Fourth International Conference on Permafrost in Fairbanks, Alaska, the International Permafrost Association was formed and now serves as a coordinating focus for international permafrost and periglacial research (Brown and Christiansen, 2005; Brown, French, and Cheng, 2008). The Ninth International Conference on Permafrost was held in Fairbanks, Alaska, in 2008, and the results published in a two-volume proceedings that included 360 peer-reviewed science and engineering papers, representing contributions from 31 countries (Kane and Hinkel, 2008; Brown, Christiansen, and Hubberten, 2008; Brown, Hubberten, and Romanovsky, 2008).

Classifications and Distribution

By definition, "permafrost," in the strict sense, refers only to the temperature of the ground; although ice may be a constituent of permafrost, this definition does not require it. Permafrost thus differs from the other three components of the cryosphere because they all consist of ice in one form or another. The limiting temperature of 0°C, however, applies to all components of the cryosphere. Thus, the existence of permafrost depends principally on *climatic* conditions. In the colder, high-latitude and high-altitude regions of the Earth (fig. 3) it is more extensive and thicker. Its distribution and behavior are sensitive to short- and long-term changes in climate. Extensive areas of subglacial, cold permafrost also occur beneath the ice sheets of Antarctica (fig. 4*A*, *B*) (Bockheim, 1995; Bockheim and Hall, 2002; Zotikov, 2006) and Greenland, as well as below many ice caps and other glaciers in the High Arctic (Humlum and others, 2005; Humlum, 2008). Large areas of subsea permafrost are also found beneath the continental shelves bordering the Arctic Ocean, especially along the Siberian coast (Rachold and others, 2007).

Permafrost ranges from very cold (temperatures of -10°C and lower) and very thick (more than 500 m to as much as 1,400 m) in the Arctic, to warm (within one or two degrees below 0°C) and thin (a few decimeters or less in thickness) in Subarctic and boreal regions (fig. 5). During glacial intervals, permafrost attained thicknesses in excess of 1,500 m in unglaciated parts of Siberia and northwestern North America. The colder and deeper permafrost is likely to have survived several interglacial periods during at least the last millions of years.

There is a long heritage of classifying and mapping permafrost at local, national (Ferrians, 1965), and international scales (Heginbottom, 2002). The permafrost region of the Northern Hemisphere is commonly divided into broad zones based on the proportion of ground area that is underlain by permafrost (fig. 3). In the northernmost zone, permafrost is found almost everywhere beneath exposed land areas; it thus forms the *continuous permafrost zone*. To the south of this zone is the *discontinuous permafrost zone*, where permafrost is absent from some areas of the landscape. This discontinuous zone is further divided into subdivisions, different terms being employed on different maps (for example, "widespread," "sporadic," "isolated patches"). In the continuous permafrost zone, permafrost occupies the entire area of

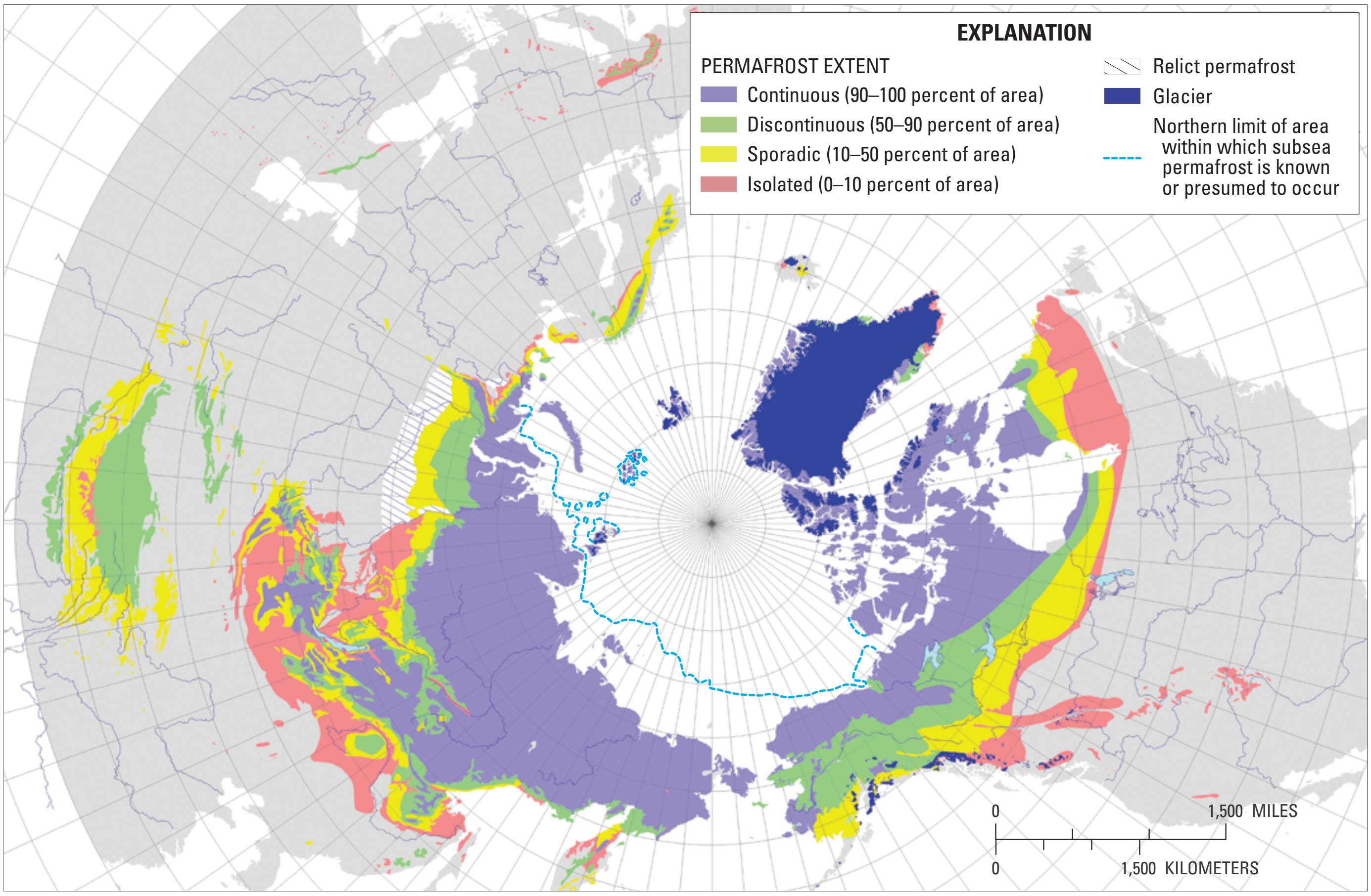


Figure 3.—*Generalized permafrost map of the Northern Hemisphere, including limit of subsea permafrost, based on the IPA Circum-Arctic Map (Brown and others, 1997; figure prepared by Dmitri Sergeev, Permafrost Laboratory, Geophysical Institute, University of Alaska Fairbanks).*

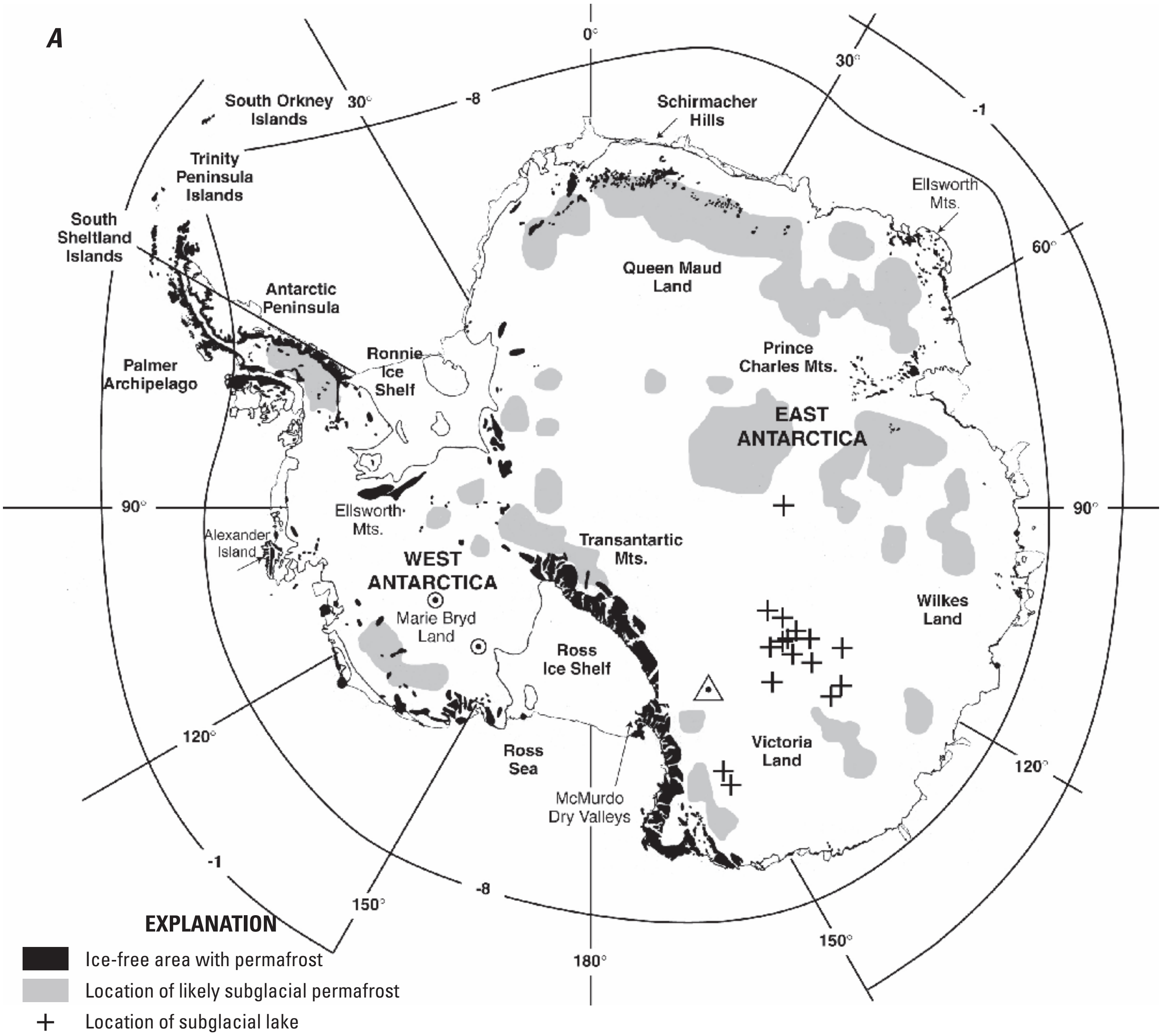
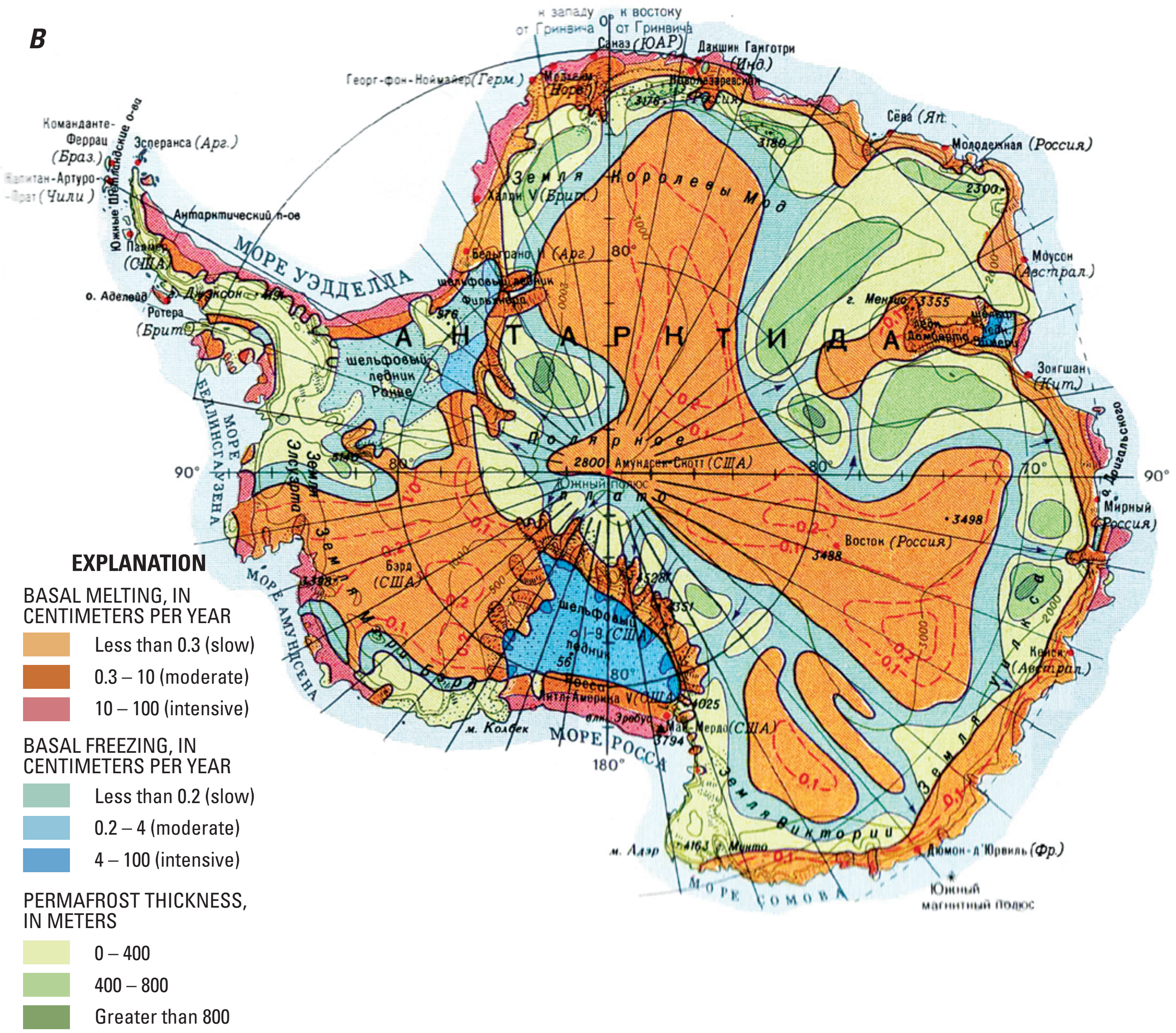


Figure 4.—*Permafrost distribution in the Antarctic:* **A**, *Generalized map (modified from Bockheim, 1995); black areas are ice-free areas with permafrost, shaded areas are location of likely subglacial permafrost, and (+) location of subglacial lakes;* **B**, *Theoretical map including subglacial permafrost distribution as originally proposed by Zotikov in 1963 with translated legend by Andrey Abramov (Kotlyakov, 1997; Zotikov, 2006).*

B



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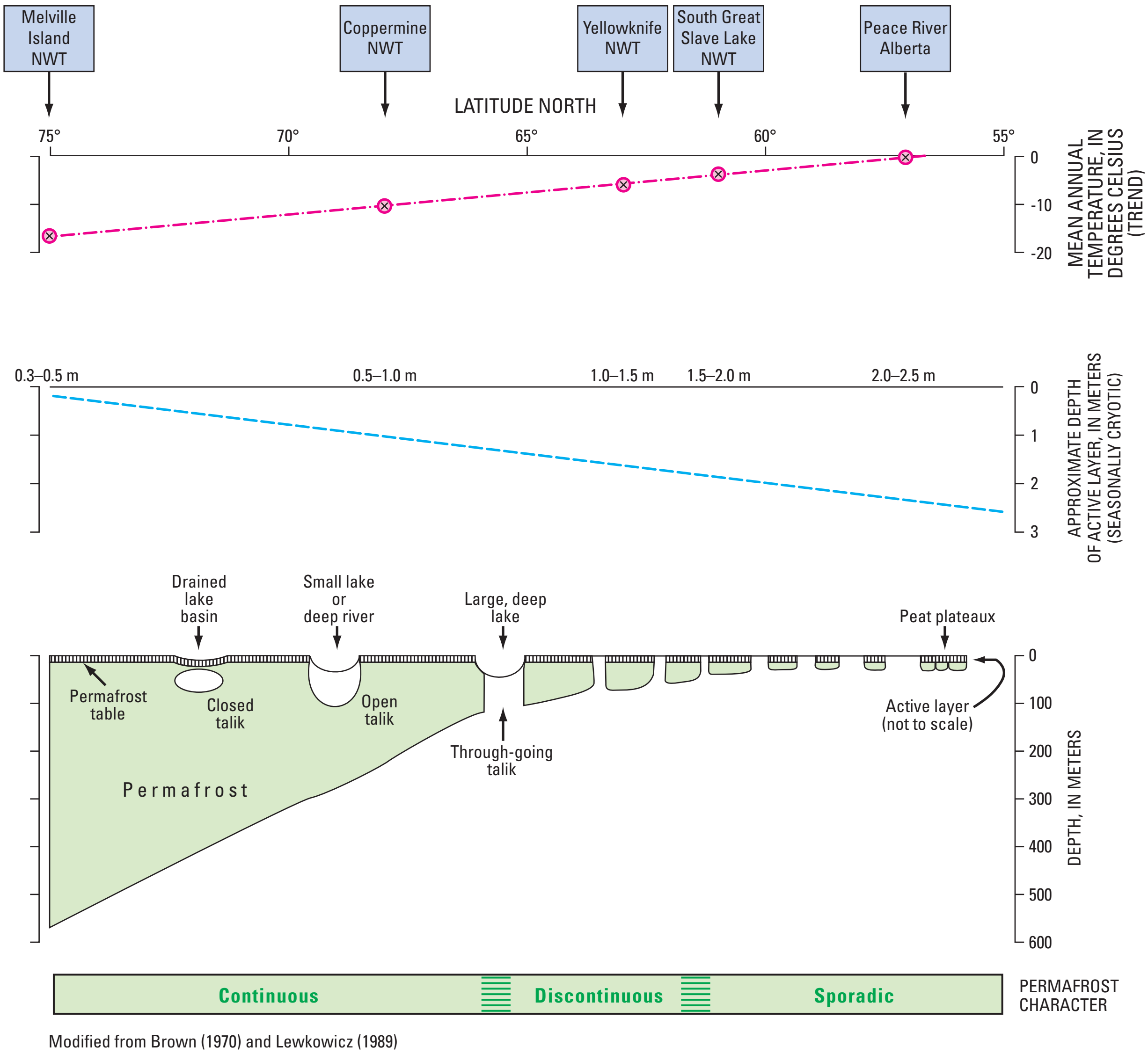

Figure 5.—*Idealized latitudinal distribution of permafrost characteristics from northwestern Canada, including the Northwest Territories (NWT) (modified from Ballantyne and Harris, 1994).*

exposed land, except beneath large rivers and deep lakes (fig. 2). In the discontinuous and sporadic permafrost zones, investigators estimate that permafrost underlies between 10 to 90 percent of the surface (table 1). Tables 2 and 3 present summaries of commonly employed classifications that recent maps have used. This variety of classifications reflects differences in criteria for continuity and for temperature. It is important that investigators be aware of these differences in classifying when they discuss the areas that permafrost occupies, especially changes in boundaries. The differing classification schemes reflect, first, increased knowledge about permafrost through time and, second, variations in the scale and purpose of the differing maps.

The delineation of the major permafrost boundaries was originally based on mean annual air temperature (MAAT) data rather than on actually measured mean annual ground temperature (MAGT) data. A MAAT of -8.5°C was used to mark the boundary between the continuous and the discontinuous permafrost zones, corresponding to a MAGT of -5°C; similarly, a MAAT -1°C was used to denote the southern boundary of the permafrost region (Brown and Péwé, 1973). In a recent analysis, Smith and Riseborough (2002) place the transition from discontinuous to continuous permafrost at a MAAT range of -6°C to -8°C.

Periglacial regions can be delimited by using one or several meteorological/climatological parameters. Jahn (1975) proposed that the -1°C isotherm of mean annual air temperature should be used to define a broad outline of the periglacial zone. In a more systematic treatment of periglacial geomorphology, French (1996, 2007) adopted the +3°C isotherm of mean annual air temperature as the boundary between periglacial and nonperiglacial conditions, because this line delimits areas of frost-induced solifluction and of patterned ground. Within the periglacial zone, French uses the -2°C isotherm for a subdivision between environments in which frost action does or does not dominate. Based on observations of sorted-ground phenomena in a periglacial environment with persistent strong winds (the Færoe Islands), Humlum and Christiansen (1998) suggested that the annual +5°C isotherm could be used to delimit windy periglacial environments, and therefore +3°C could be used

Table 1.—*Area of individual permafrost classes within the exposed land surface of the Northern Hemisphere*

[Based on the International Permafrost Association (IPA) permafrost classification method using continuity (extent) and ground ice content (Brown and others, 1997; Brown and Haggerty, 1998) and on the IPA Circum-Arctic map (Brown and others, 1997; Zhang and others, 1999. 10^6 km², million square kilometers; m, meters; <, less than; >, greater than; %, percent)]

	Ice content	Lowlands/uplands with thick overburden (>5–10 m)			Mountains with thin overburden (<5–10 m)		Total (10^6 km²)
		High >20%	Medium 10–20%	Low 0–10%	High >10%	Low 0–10%	
Extent of permafrost	Continuous (>90%)	1.42	1.27	0.37	2.09	5.54	10.69
	Discontinuous (50–90%)	0.08	0.85	0.38	0.75	2.32	4.38
	Sporadic (10–50%)	0.11	0.30	0.56	0.32	2.61	3.90
	Isolated (0–10%)	0.34	0.07	0.58	0.03	2.80	3.82
	Total areas (10^6 km²)	1.95	2.49	1.89	3.19	13.27	22.79

Table 2.—*Different limits on the continuity of permafrost based on the percentage of the land area underlain by permafrost*

[From Heginbottom (2002, table 1); <, less than; >, greater than; ?, undefined]

<table>
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(5–40)</td><td colspan="16">Discontinuous permafrost
(40–80)</td><td colspan="9">Continuous permafrost
(80–100)</td></tr>
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(10–25)</td><td colspan="10">Insular occurrence
(25–50)</td><td colspan="12">Massive insular occurrence
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(80–100)</td></tr>
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(80–100)</td></tr>
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(70–90)</td><td colspan="5"></td></tr>
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(10–35)</td><td colspan="12">Intermediate discontinuous permafrost
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(10–50)</td><td colspan="16">Extensive discontinuous permafrost
(50–90)</td><td colspan="5">Continuous permafrost
(>90)</td></tr>
<tr><td colspan="43">International Permafrost Association</td></tr>
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Brown and others (1997)</td><td colspan="5">Isolated patches
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(10–50)</td><td colspan="16">Extensive discontinuous permafrost
(50–90)</td><td colspan="5">Continuous permafrost
(>90)</td></tr>
<tr><td>van Everdingen (1998)[4]</td><td colspan="3">?</td><td colspan="10">Sporadic permafrost
(5–30)</td><td colspan="4">?</td><td colspan="8">Island permafrost
(40–60)</td><td colspan="4">?</td><td colspan="4">Massive island permafrost
(70–80)</td><td colspan="9">[Continuous permafrost]
(80–100)</td></tr>
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</table>

[1]From Trofimov and Kondratyeva (1989, figure 33)

[2]From translations and commentary, Williams and Warren (1999)

[3]Predominantly unfrozen ground

[4]From International Permafrost Association Glossary (Russian usage), based on Kudryavtsev, et al (1978)

[5]From International Permafrost Association Glossary (English usage), based on Heginbottom and Radburn (1992)

Table 3.—Different limits on the continuity of permafrost based on permafrost temperature

[<, less than, >, greater than]

<table>
<tr><th rowspan="2">Reference</th><th colspan="20">Permafrost temperature, in degrees Celsius</th></tr>
<tr><th>+10.0</th><th>+5.0</th><th>+4.0</th><th>+3.0</th><th>+2.0</th><th>+1.0</th><th>+0.5</th><th>0</th><th>-0.5</th><th>-1.0</th><th>-1.5</th><th>-2.0</th><th>-3.0</th><th>-4.0</th><th>-5.0</th><th>-7.0</th><th>-9.0</th><th>-11.0</th><th>-13.0</th><th><-13.0</th></tr>
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<tr><td colspan="7"></td><td colspan="2">Scattered islands</td><td colspan="2">Isolated patches</td><td colspan="9"></td></tr>
<tr><td colspan="4"></td><td colspan="11">Discontinuous permafrost</td><td colspan="5"></td></tr>
<tr><td rowspan="5">Gravis and others (1973), at ground surface</td><td colspan="6">Seasonally frozen ground</td><td colspan="14"></td></tr>
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<tr><td rowspan="7">Yershov and others (1999)</td><td colspan="4">Continuously unfrozen</td><td colspan="16"></td></tr>
<tr><td colspan="3"></td><td colspan="3">Predominantly unfrozen</td><td colspan="14"></td></tr>
<tr><td colspan="4"></td><td colspan="5">Scattered islands</td><td colspan="11"></td></tr>
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<tr><td colspan="12"></td><td colspan="8">Continuous permafrost</td></tr>
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as a warm limit for periglacial environments with little wind action. From a geographical point of view, it is of great interest to map periglacial features in order to support palaeoenvironmental evaluations and the reconstruction of the cold climate environments during the Quaternary Period glaciations (Vandenberghe, 2001).

Permafrost in lower-latitude mountainous regions and plateaus is commonly referred to as *mountain* (or *alpine*) *permafrost* and *plateau permafrost*, respectively (King, 1983; Qiu and others, 2000). In mountainous terrain, the continuity of permafrost is difficult to assess, because strong environmental, ecological, and microclimatological factors set limits on elevation, slope, aspect, vegetation, and snow cover. There are also difficulties in portraying the continuity of permafrost zonation at these scales (fig. 6). Modeling approaches provide more flexible methods of portraying the probable or possible occurrences of permafrost or their lower limits, as illustrated in northern Mongolia (Etzelmüller and others, 2006). On the Qinghai-Xizang (Tibet) Plateau, broad zones of predominantly continuous and isolated permafrost may be recognized in addition to mountain permafrost.

In the early 1990s, the lack of a global map of permafrost with a common legend presented problems for assessing climate change. In response to this deficiency, the International Permafrost Association (IPA) prepared and published a unified map of the Northern Hemisphere at a scale of 1:10,000,000 (Brown and others, 1997; Heginbottom and others, 1993), using a two-parameter classification system based on continuity of permafrost and on ground-ice content (fig. 7). The mapping principles were based in part on the previously published Canadian map (Heginbottom and others, 1995). A digital version of the IPA map enabled investigations to develop a statistical evaluation of permafrost distribution across the entire Northern Hemisphere, calculating that the permafrost regions occupy up to 22.79×10^6 km^2 or 23.9 percent of the exposed land area of the Northern Hemisphere (table 1) (Brown and Haggerty, 1998; Zhang and others, 1999) and that glaciers in the Northern Hemisphere occupy an area of 2.12×10^6 km^2. Because permafrost is not present everywhere in the permafrost region, the actual land area that it underlies is calculated to range

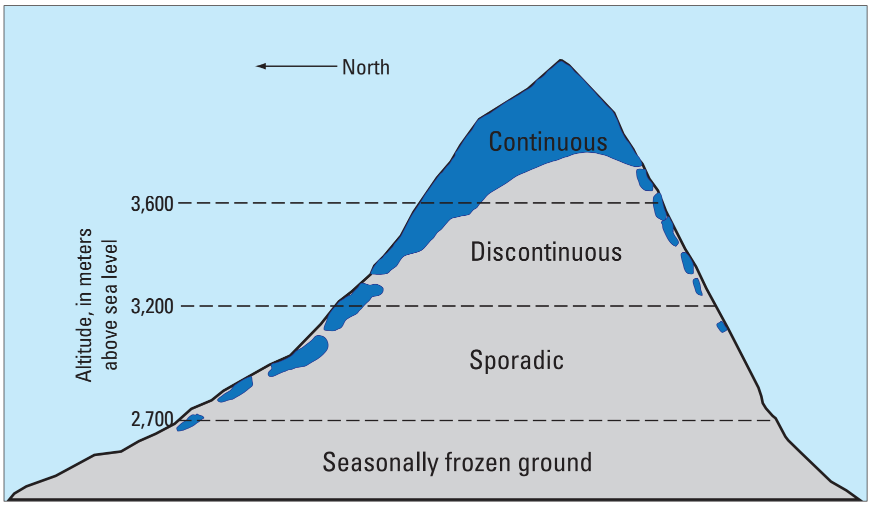


Figure 6.*—Idealized diagram of altitudinal distribution of sporadic, discontinuous, and continuous permafrost (figure provided by Sergei Marchenko, Permafrost Laboratory, Geophysical Institute, University of Alaska Fairbanks).*

Extent of permafrost (percent of area)	Ground-ice content (Visible ice in the upper 10–20 m of the ground, in percent volume)				
	Lowlands, highlands, and intra- and intermontane depressions characterized by thick, overburden cover (f) (>5–10 m)			Mountains, highlands, ridges, and plateaus characterized by thin, overburden cover (<5–10 m) and exposed bedrock (r)	
	High (>20 percent)	Medium (10–20 percent)	Low (0–10 percent)	High to Medium (>10 percent)	Low (0–10 percent)
Continuous (90–100 percent)	Chf (1)*	Cmf (5)	Clf (9)	Chr (13)	Clr (17)
Discontinuous (50–90 percent)	Dhf (2)	Dmf (6)	Dlf (10)	Dhr (14)	Dlr (18)
Sporadic (10–50 percent)	Shf (3)	Smf (7)	Slf (11)	Shr (15)	Slr (19)
Isolated patches (0–10 percent)	Ihf (4)	Imf (8)	Ilf (12)	Ihr (16)	Ilr (20)

*Letters and numbers adjacent to color codes refer to polygon classes in Arc/Info database.

Figure 7.*—Legend for the Circum-Arctic map of permafrost and ground-ice conditions of the Northern Hemisphere (Brown and others, 1997).*

from 12.2 to 16.9×10^6 km^2 or from 13 to 18 percent of the exposed land area (Zhang and others, 1999, 2000). These computations exclude bodies of thawed ground (taliks) existing beneath deep lakes in the continuous permafrost zone, and they represent the maximum and minimum estimates of the area underlain by permafrost in the Northern Hemisphere. The regions underlain by permafrost therefore comprise about 60 percent of present Russia, 50 percent of Canada, 23 percent of China, and 90 percent of Alaska.

Compared to knowledge about the Northern Hemisphere, relatively little is known about the distribution, thickness, and properties of permafrost in the Southern Hemisphere. It is known that permafrost occurs at high elevations throughout the length of the Andes of South America (from Ecuador, through Bolivia and Perú, to Argentina and Chile), in the subantarctic islands of the Southern Ocean, and in Antarctica (fig. 4*A*). Based on a limiting mean annual air temperature isotherm of -5°C, the area of permafrost in South America is estimated to be 100 to 150×10^3 km^2 (fig. 8). In the central Andes, the possible presence of permafrost within the periglacial area extends over about 17×10^3 km^2, although the area actually glacierized covers only 1.26×10^3 km^2 (Trombotto, 2000; D. Trombotto, Department of Geocryology, Instituto Argentino de Nivología Glaciología y Ciencias, Mendoza, Argentina, written commun., January 2006). Many areas require more detailed investigations, especially areas recently deglacierized as mountain glaciers, ice caps, ice fields, and their associated outlet glaciers recede.

Bockheim (1995) published the map of permafrost in the Antarctic region (fig. 4*A*), indicating that permafrost existed throughout the ice-free areas (an estimated 49×10^3 km^2) and with subglacial permafrost occurring only in areas

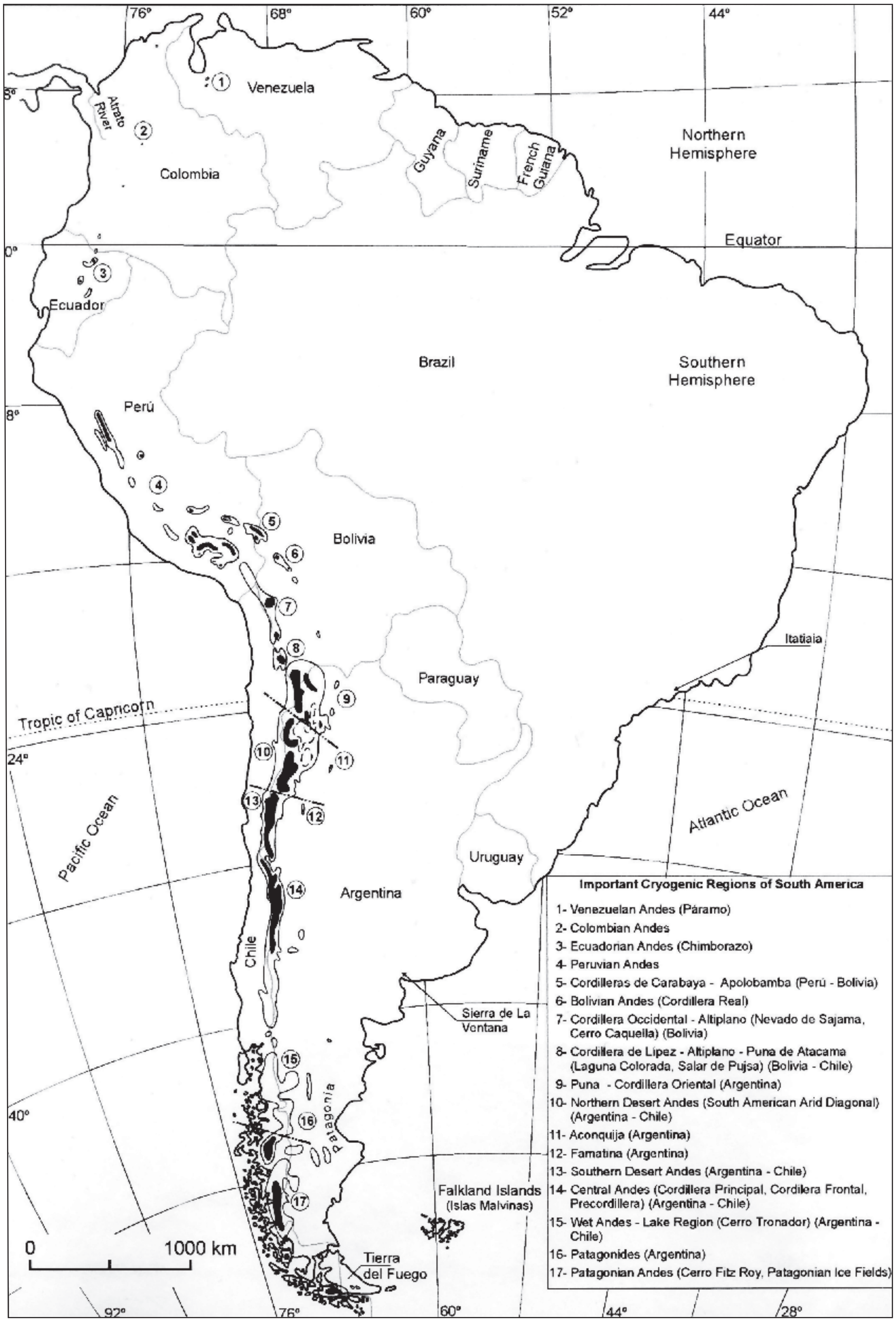


Figure 8.—*Cryogenic regions of South America; black is area of mean annual air temperature (MAAT) of 0° to -5°C, including ice covers (after Trombotto, 2000).*

where the ice sheet was thin enough to maintain frozen subglacial conditions (Herterich, 1988). Bockheim and others (2007) distinguish three types of permafrost in the McMurdo Dry Valleys and elsewhere on the continent: dry-frozen, ice-cemented, and ground-ice or buried ice. In the subantarctic islands of the Southern Ocean, permafrost occurs at higher elevations.

Understanding the distribution of permafrost in continental Antarctica is complicated by the thickness of the ice sheet in East Antarctica, which causes pressure melting at the base, resulting in the presence of large lakes (Oswald and Robin, 1973; fig. 4*A*, *B*). The existence of 145 known lakes beneath the Antarctic ice sheet has since been confirmed from airborne radar profiling (Kennicutt and Petit, 2007). Most such lakes are about 10 km in length, but the largest, Lake Vostock, is about 250 km in length.

Thermal Regimes

The occurrence and distribution of permafrost depend on the heat balance at the Earth's surface. When the depth of seasonal freezing exceeds the depth of seasonal thaw of the soil, permafrost begins to grow downward at the base of seasonal frost. The lower limit of permafrost eventually reaches an equilibrium depth when the heat flow from the interior of the Earth (local and regional geothermal gradient) offsets the surface temperature. The permafrost temperature regime (at depths of 10 to 200 m) is a sensitive indicator of decade-to-century climatic variability and of long-term changes in the surface-energy balance. The range of variations in interannual temperature ("noise") decreases significantly with depth, but variations in decadal and longer time scales (the "signal") are found at greater depths in permafrost with less attenuation. As a result, the "signal-to-noise ratio" increases rapidly with depth, and the ground acts as a natural low-pass filter of the climatic signal. This relation of temperature-depth profiles in permafrost provides a useful technique for studying past temperature changes at the ground surface (Lachenbruch, 2002; Romanovsky and others, 2002; Osterkamp, 2003b). For example, from borehole temperature measurements in abandoned oil wells in northern Alaska, Lachenbruch and Marshall (1986) detected a 20th-century warming in permafrost that began in the late 19th century.

Permafrost temperatures respond to changes in surface climate, hydrology, and vegetation, to landscape changes, such as those caused by erosion (for example, lake drainage or marine transgressions) and wildfire, and to human-induced changes including disruption of the surface or change in surface cover during construction. Seasonal temperature oscillations are generally less than 0.1°C at depths below 15 to 20 m, and summer and winter temperature profiles intersect at this depth as annual temperature rises negligibly. Examples of temperature profiles spanning four decades from geographically diverse regions of the Northern Hemisphere are shown in figures 9*A*–*F* (Brown, Hubberten, Romanovsky, 2008). Field observations since the 1980s from an array of shallow boreholes in northern Alaska demonstrate a gradual warming of permafrost temperatures at this zero amplitude depth (fig. 9*D*). The monitoring of a large number of existing and new boreholes occurred as part of the International Polar Year (IPY) and its Thermal State of Permafrost (TSP) Project, details of which are discussed in the section "Observational Networks" below.

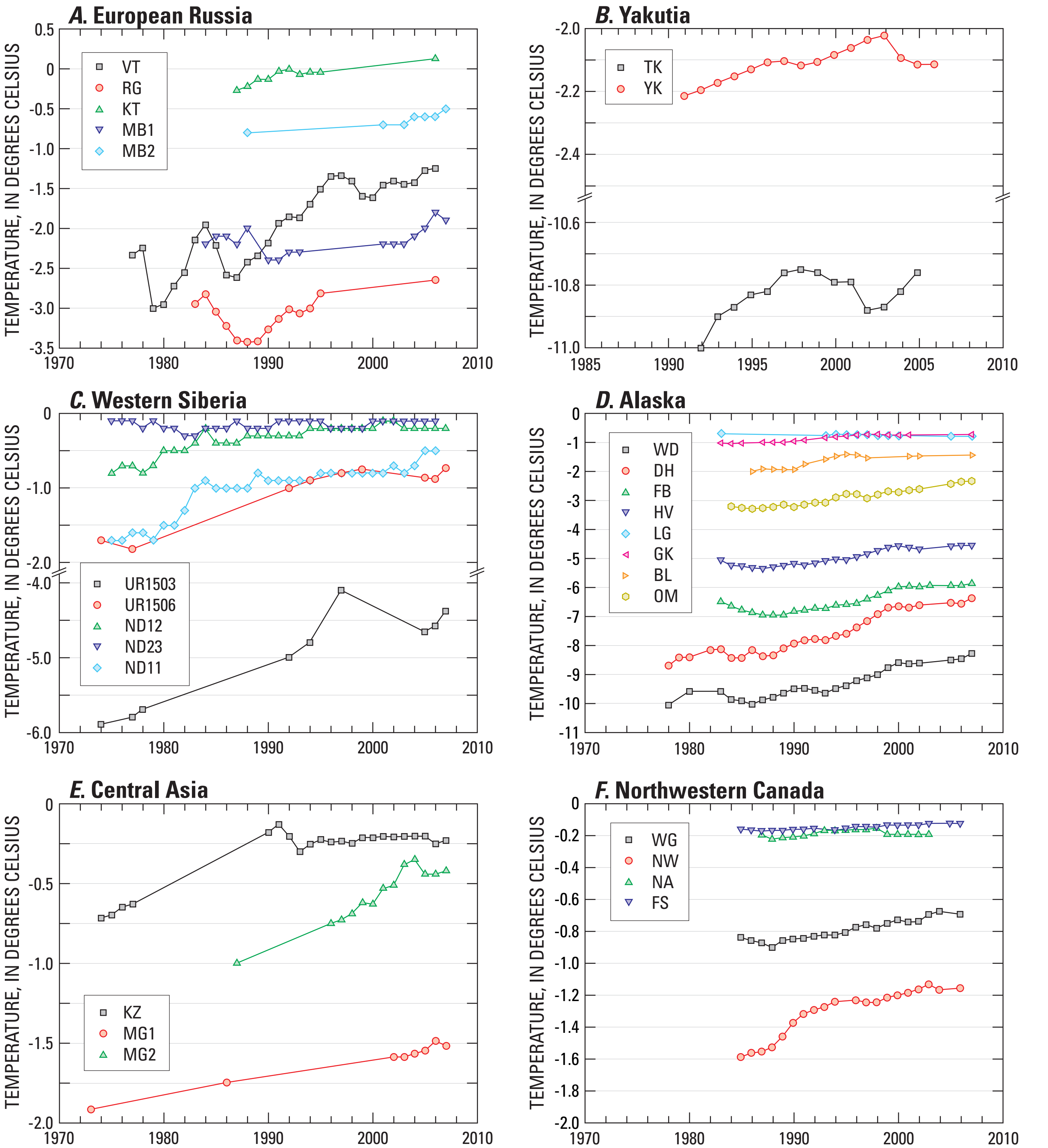


Figure 9.—*Long-term trends in permafrost temperatures for selected locations in the Northern Hemisphere (modified from Brown, Hubberten, and Romanovsky, 2008).* **A**, *Permafrost temperatures from European Russia (VT-Vorkuta; RG-Rogovoi; KT-Karataikha; MB-Mys Bolvansky);* **B**, *Permafrost temperatures from Yakutia, Russia (TK-Tiksi; YK-Yakutsk);* **C**, *Permafrost temperatures from western Siberia (UR-Urengoi; ND-Nadym);* **D**, *Changes in permafrost temperatures at 20-m depth in Alaska (WD-West Dock; DH-Deadhorse; FB-Franklin Bluffs; HV-Happy Valley; LG-Livengood; GK-Gulkana; BL-Birch Lake; OM-Old Man);* **E**, *Permafrost temperatures from Central Asia (KZ-Kazakhstan; MG-Mongolia); and* **F**, *Ground temperatures at depths of 10 to 12 m between 1984 and 2006 (A–F) northwestern Canada (WG-Wrigley; NW-Norman Wells; NA-Northern Alberta; FS-Fort Simpson).*

Under the current phase of climatic warming, it has been observed locally that the spatial distribution of warmer permafrost is decreasing (Arctic Climate Impact Assessment [ACIA], 2004, 2005; Jorgenson and others, 2001; Camill, 2005). Model results based on scenarios for 21st-century climate change forecast increased thawing of permafrost and a shift of contemporary ground-temperature zones northwards and to higher elevations in mountain ranges (Arctic Climate Impact Assessment [ACIA], 2004, 2005; Romanovsky, Armstrong, and others, 2007). Although some shifts in permafrost boundaries may occur, hundreds (perhaps even thousands) of years are required to substantially degrade thick deposits of ice-rich permafrost and to significantly increase the areal extent of open taliks. In addition, regional variations in geothermal heat flow will influence thawing of permafrost from below.

Permafrost and ground ice are still forming in the 21st century along aggrading and rebounding shorelines (such as islands in the Canadian High Arctic and around the shores of Hudson Bay), in the exposed beds of newly drained lakes (Mackay, 1998) within the colder parts of the permafrost region, and along the foot of many mountain slopes with talus accumulation and avalanche activity (Humlum, 2005).

In high northern latitudes, permafrost probably developed during the Pliocene Epoch (-3.5 Ma) (Lunardini, 1995; Osterkamp, 2003b) and may be older in Antarctica. During the numerous glacial intervals of the Pleistocene Epoch, when global sea levels were as much as 125 m lower than they were in 2008 (Fairbanks, 1989), large areas of the Arctic continental shelves of northern Russia and North America were exposed to the full glacial climate of subzero °C temperatures (Romanovskii and Hubberten, 2001). In those areas now beneath the sea floor, permafrost developed to thicknesses in excess of 700 m, and relict permafrost still survives beneath the sea floor. In areas that were covered by the Pleistocene ice sheets, thermal models suggest that preexisting permafrost and ground ice would have disappeared from beneath warm sectors at the base of the ice sheets (Humlum and others, 2003). Thus, in areas that were glaciated, much of the permafrost and ground ice has developed in postglacial time (during the late Pleistocene and Holocene Epochs). In unglaciated parts of northern Russia, Alaska, and Yukon, permafrost and ground ice may be much older. In northwestern Canada, for example, some ground ice is believed to predate the last glaciation (Mackay and others, 1972), and, in Svalbard, permafrost in the high mountains is thought to be perhaps as much as 800,000 years old or even older (Humlum and others, 2003). Permafrost also developed in periglacial environments around the southern margins of the continental ice sheets in the Northern Hemisphere and also on the eastern margin of the Patagonian ice sheet in southern South America, as evidenced by the widespread occurrence of a range of diagnostic landforms, such as casts of former ice wedges and circular ridges formed by the collapse of pingos. Heginbottom (2004) reviewed changes in permafrost distribution over geological time scales and over recent decades.

Ground Ice: A Component of the Cryosphere

Ground ice is a general term for all types of naturally occurring ice formed or found in freezing and frozen ground. Although it occurs in a wide variety of forms as a result of diverse processes and is common in many permafrost, periglacial, and seasonal freezing environments, only three main classes are recognized: pore ice, ice lenses and veins, and larger bodies of more or less pure ice. Pore ice, when thawed, will not exceed the natural void volume of the sediments. Excess ice, when thawed, yields water that exceeds the void volume; this water will accumulate at the surface as the newly thawed ground consolidates and subsides. Sediments that contain substantial amounts of excess ice are termed "ice rich." Perennial ground ice, including buried, subsurface ice, cannot exist independently of permafrost. Freezing and thawing processes and the presence of excess ice are responsible for the many landscape features associated with ground ice.

In permafrost regions, ground ice is one of the most important attributes of the terrain (Mackay, 1972). Its presence influences topography, geomorphic processes, vegetation, and the landscape's response to environmental changes, whether natural or technogenic. Most perennial ground ice forms in place either by segregation or injection. Segregated ice forms as a freezing front advances from the surface into the unfrozen ground beneath. Pore water within the unfrozen ground is drawn to the freezing front by suction, where it accumulates as thin lenses, commonly alternating with layers of soil. Injection ice, or intrusive ice, forms by the freezing of water injected into soils or rocks. Ground ice may form more or less simultaneously with the enclosing sediments, as *syngenetic* ice; or it may form later, as *epigenetic* ice. In areas of seasonal frost, however, ground ice can also have important effects on geomorphic processes and human activities even though it is ephemeral. It occurs primarily as lenses and infillings of minor frost cracks a few millimeters to centimeters thick. Ice-lens formation causes frost heaving of the ground surface, with subsequent weakening by thaw. The combination can damage roads and other structures whose foundations or bases do not extend deeper than the boundary of the subsurface freeze.

In contemporary or recently glacierized regions, ground ice also includes ice in ice-cored moraines and other buried masses of glacier ice. In formerly glaciated areas, buried ice-sheet ice may form an extensive component of extant ground ice. Buried surface ice may have started as river or lake ice, sea ice, glacier ice, late snow beds, or the ice from surface icings (naledi or aufeis) and, having been buried rapidly, became incorporated into permafrost. Burial mechanisms include landslides, slope wash, river sedimentation, eolian activity, deposition of moraines, and glacial outwash sands and gravels.

The amounts of ground ice present have been estimated globally and in selected areas. For the Northern Hemisphere, the amount of ground ice in permafrost (in the uppermost 10 to 20 m of the ground) is estimated to range from 10.8 to 35.5×10^3 km^3; the volume of ice in seasonally frozen ground is assumed to be considerably less (Brown and Haggerty, 1998; Zhang and others, 1999, 2000). The Intergovernmental Panel on Climate Change (IPCC), in its Second Assessment Report (Fitzharris and others, 1996), stated that the modern volume of ice in permafrost globally is 160×10^3 km^3. Ground ice is generally more abundant near the surface, then decreases with depth; most ground ice in permafrost is found within 10 m of the ground surface; locally, estimates of nearly 45 percent by volume of the upper several meters of permafrost have been reported for northern Alaska and Richards Island, Canada

(Pollard and French, 1980). In the Russian Arctic, the volumetric ice content of peat and unconsolidated sediments approaches 80 percent in lacustrine and organic-lacustrine deposits (E.S. Melnikov, written commun., Earth Cryosphere Institute, Moscow, Russia, 1992), particularly in the "*edoma*" deposits of eastern Siberia where the ice complex is as much as 40 m in thickness (Sher and others, 2005).

Detailed examination of ice-bearing sediments can provide information on the nature of the freezing process and of the conditions under which the sediments accumulated. Distinct sedimentary structures (termed "cryostructures") develop as a result of the freezing process, depending on the initial water content and the extent of water migration during freezing. Cryostructures may be grouped into cryofacies or analyzed in terms of their cryotextures (Murton and French, 1994; French, 1996; Shur and Jorgenson, 1998) (fig. 10).

Pore Ice, Ice Lenses, and Veins

Pore ice includes individual ice crystals within the pores of sedimentary rocks and unconsolidated deposits and icy coatings on individual soil particles. On melting, pore ice does not yield water in excess of the void volume of the same soils or rock when unfrozen. If sufficient pore ice is present, it acts as cement, bonding the soil particles together, and greatly increases the strength of the ground. Most ground ice occurs in the form of thin, sub-horizontal sheets with tapering edges—*ice lenses*. These vary in thickness from a hairline thickness to more than 10 m; very thick and extensive ice lenses are better termed "massive-ice beds" (see the separate discussion). The term "ice lens" is commonly used for a layer of segregated ice that is more or less parallel to the ground surface. Ice lenses are frequently located at changes in soil texture. Ice veins are similar to lenses, but they cut across stratigraphic and soil horizons; vein ice develops in tension or dilation cracks or as a result of water intrusion. In soils with a high clay content, the freezing process may produce a three-dimensional network of veins of relatively pure ice surrounding soil blocks containing little ice, a form that is called "reticulate vein ice."

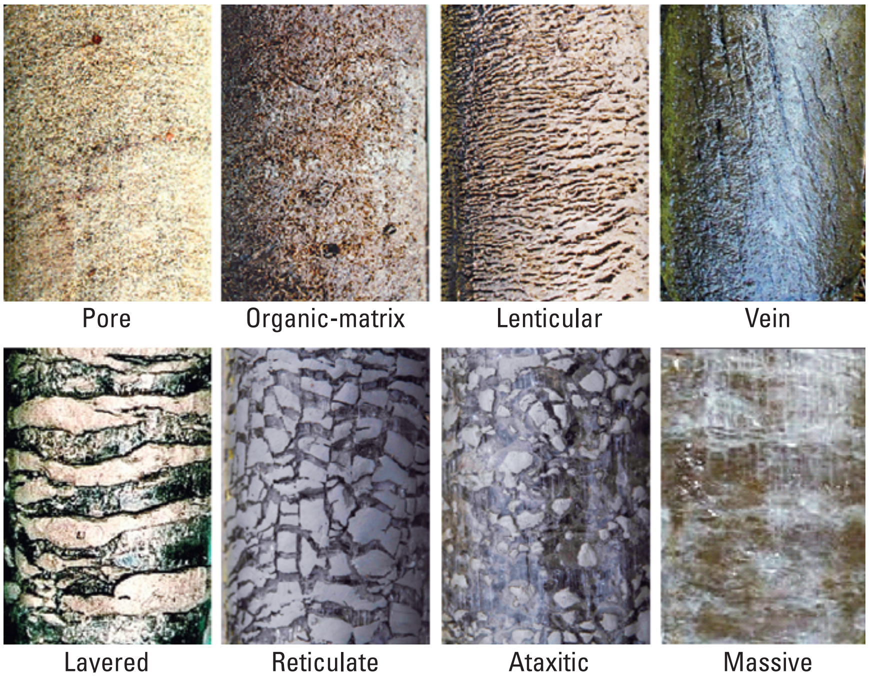


Figure 10.—*Illustrations of eight types of segregated ice in permafrost cores (provided by M.T. Jorgenson, ABR, Inc., Fairbanks, Alaska).*

Ice-Wedge Ice

Ice wedges consist of vein ice and typically take the form of vertical wedge-shaped dykes penetrating downwards into the permafrost (fig. 11) (Péwé, 1966). They vary from less than 10 cm to more than 3 m in width at the top, commonly tapering to a feather edge at depths of 1 m to greater than 10 m; in syngenetic deposits, some ice wedges may extend as deep as 25 m. Lachenbruch (1962) provided a detailed analysis of the mechanism of ice-wedge formation (fig. 12), based on the contraction theory first proposed by Leffingwell (1919). Cracking results in a polygonal network of vertical cracks that relieve tensile stress set up by such contraction. Meltwater percolates into the cracks during spring and then refreezes, preventing the cracks from closing fully as the permafrost warms and expands in summer. Because the tensile strength of ice is less than that of frozen sediment, thermal-contraction cracks tend to reopen along the preexisting network of ice veins in subsequent winters

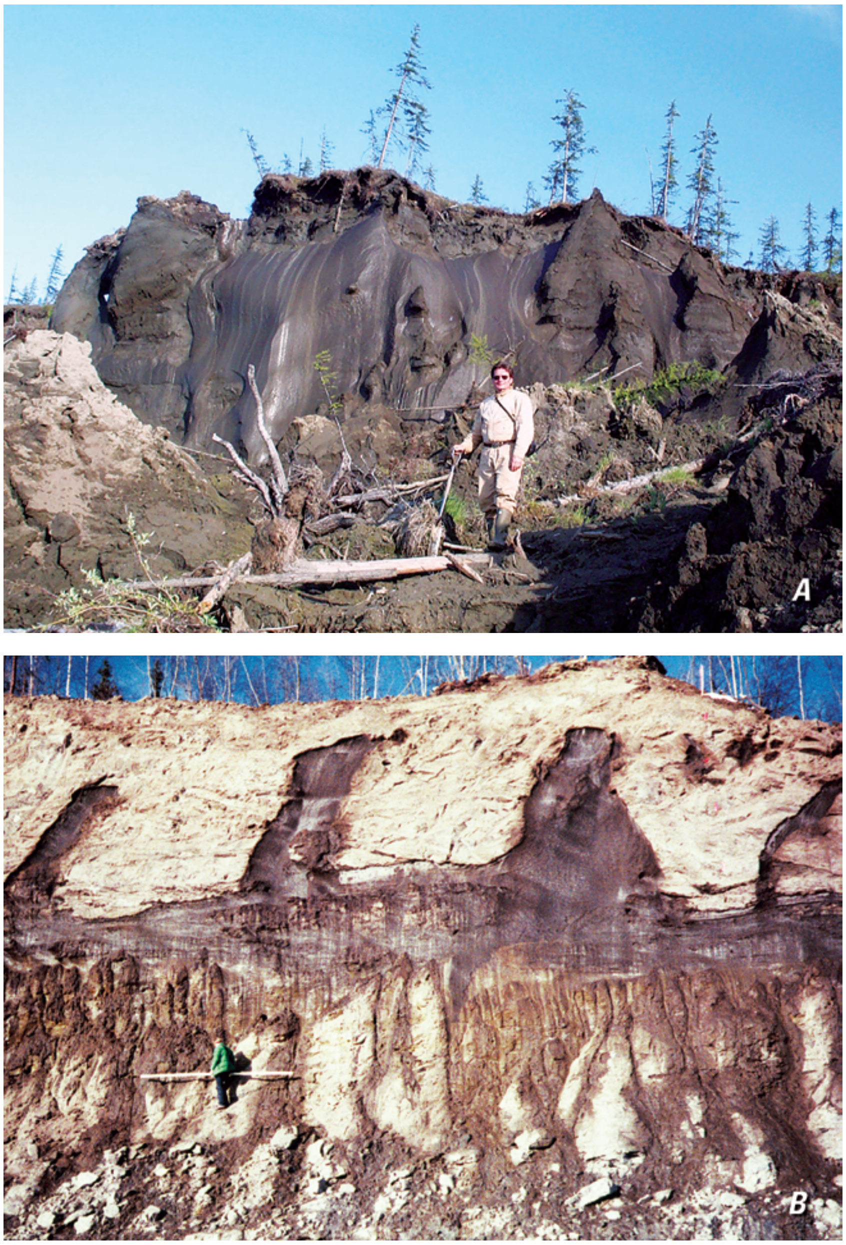


***Figure 11.**—Two examples of large ice wedges:* **A**, *Massive ice wedges and thawing of permafrost along the bank of the Kolyma River, Siberia, Russia (photograph provided by Vladimir Romanovsky); and* **B**, *Ice wedges exposed in road cut along Steese Highway near Fox (Fairbanks), Alaska. (1977 photograph provided by Steven Arcone, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory).*

(Mackay, 1992). Ice wedges thereby grow incrementally as each successive phase of contraction cracking is accompanied by filling of the open crack with more ice. The incremental growth of wedge ice produces foliation parallel to the sides of the wedge. Such foliation is due to aligned ice crystal fabrics and the incorporation of air bubbles and silt layers.

Periods of warmer climate thaw the ground more deeply, leading to truncation of the upper surfaces of ice wedges. A subsequent cooler period, with a thinner active layer, may reactivate growth of the ice wedge, with a small secondary wedge nested in a deeper, larger, older wedge. In eastern Siberia, ice in the complex of the *edoma* deposits can occupy the upper 30 to 40 m of icy silts that comprise a network of large syngenetic ice wedges (figs. 13*A*, *B*; Sher and others, 2005). Radiocarbon dating of organic remains in ice wedges, pollen analyses, and the isotopic composition of the ice provide evidence for the Pleistocene age of these deposits (Vasil'chuk and others, 2004).

Massive Ice

Thick and extensive ice lenses are termed "massive-ice beds." They can be up to tens of meters in thickness and hundreds of meters in horizontal extent. In North America and Russia, such ice forms are widespread in the lowlands of the Arctic coastal region. The origin of massive ice has been the cause of much debate; early workers in Russia concluded that massive ice bodies seen in river banks were large ice wedges exposed along their length. More recently, North American scientists have proposed that massive ice bodies in northwestern Canada have originated in situ by ice-segregation processes, possibly aided by injection of water under pressure generated by the waning Wisconsinan ice sheet (Mackay, 1971). Other workers, particularly in Russia, have suggested that most massive ice has its origin in the burial of surface ice (for example, from remnants of glacier ice). Although isotope studies from northern Canada provide increasing evidence today (2012) that some massive ice is indeed buried glacier ice, a segregation origin is still the favored explanation for most such ice masses.

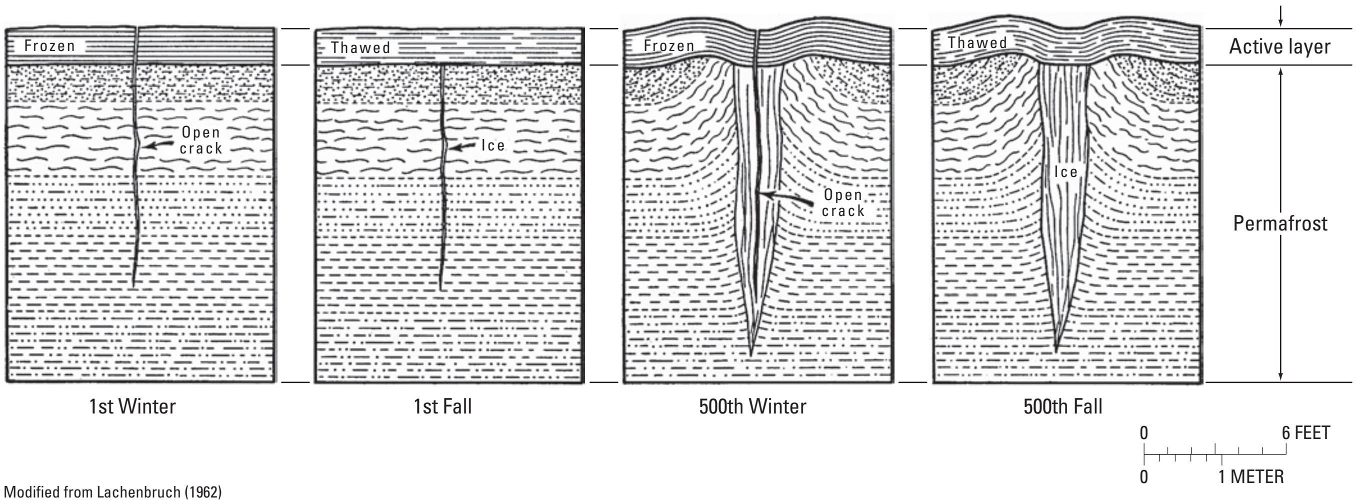
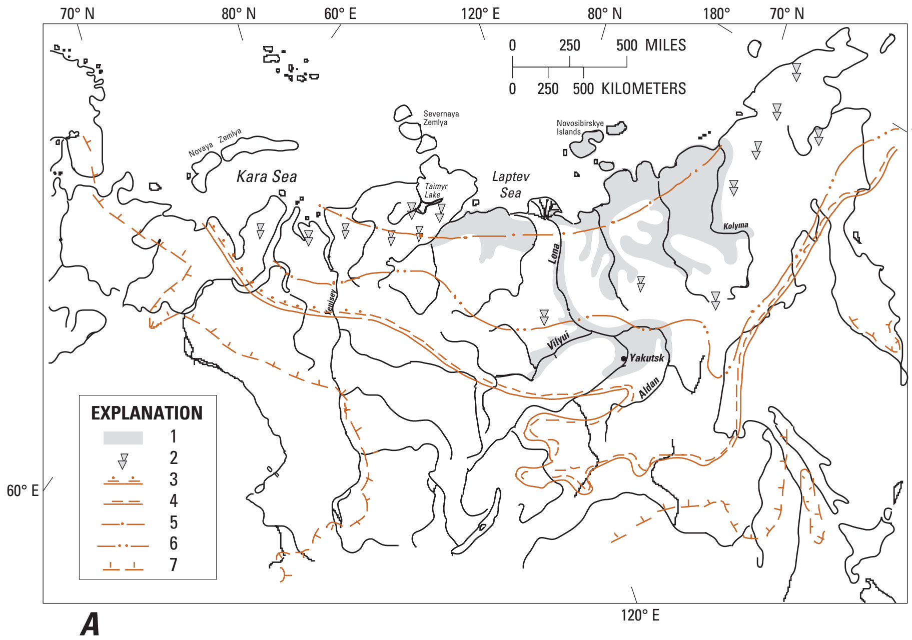


Figure 12.*—Schematic drawing of the evolution of an ice wedge according to the contraction-crack theory (Lachenbruch, 1962).*

Needle Ice

Needle ice (also known by the Swedish term *pipkrake*; Neuendorf and others, 2005, p. 494) is a diurnal form of shallow, subsurface ice that consists of groups of narrow ice slivers, which can be up to several centimeters long. The groups of slivers normally form in moist soils when temperatures drop below freezing during the night. Needle ice plays an active role in loosening soil for erosion, and it tends to move small rocks upward to the soil surface. On sloping surfaces, needle ice can also enhance soil creep by moving soil particles at right angles to the grade, a process that can be termed "needle-ice creep." The formation of needle ice during nights in areas where there is strong radiational cooling (no cloud cover) can lead to extensive disruption of the ground surface and to areas of locally rapid solifluction (Mackay and Matthews, 1974; Washburn, 1980).



A

B

Figure 13.—**A**, *Schematic map of the Edoma "Ice Complex" in northern Siberia. (1) Areas of extensive occurrences of the ice complex across a range of landforms; (2) areas of the ice complex occurring only in river valleys and lake depressions; 3–5 southern boundaries of regions of contemporary ice wedge development: in (3) peat deposits; (4) silt and clay deposits; (5) sand and gravel deposits; (6) southern boundary of the regions of low-centered polygons; (7) southern boundary of permafrost (N.N. Romanovskii, 1993).* **B**, *Exposure of eroding Edoma deposit along the Laptev Sea coast, northern Siberia, Russia (photograph by Volker Rachold).*

Landforms Dominated by Ground Ice

In the regions dominated by permafrost and periglacial processes, the underlying ground ice is a primary component of the landscape. The resulting permafrost landforms are periglacial by nature and in their geographical distribution. Extensive field observations throughout the Arctic and Subarctic attest to the widespread occurrence of massive ground ice, ice-cored terrain, and soils with high ice content. Ground ice can produce 50 to 75 percent of the local topographic relief, and it can constitute as much as 50 to 70 percent by volume of the upper 5 to 10 m of the permafrost over broad areas (Brown and others, 1997).

Ice-Wedge Polygons

A striking feature of many flat regions (for example, plains, terraces, valley floors, and even gently sloping sides of valleys) is the presence of polygonal networks formed by shallow troughs in the ground surface (fig. 14). The surface troughs mark the locations of underlying ice wedges that commonly extend 3 to 5 m downwards into the permafrost from the base of the active layer. The polygons generally range from 15 to 50 m in diameter. Their form may be roughly hexagonal, with angles tending to approximate 120°, or orthogonal with a tendency towards 90° junctions (Lachenbruch, 1962). The typical surface expression of an active ice wedge is a shallow trough, bounded by a pair of low ridges. The width of the trough is indicative of the minimum width of the ice wedge, and the ridges result from deformation of the sediments by the growing ice wedge. On flat terrain, low-centered polygons frequently form shallow ponds. Polygon fields are among the most widespread and visible permafrost features of the landscape of arctic regions.

Actively growing polygons occur primarily in areas of cold, continuous permafrost. Inactive polygons, which often have no polygonal surface expression, can be stable and remain for many years to centuries without changing; they occur in the continuous as well as the discontinuous zones of permafrost.

Figure 14.*—Oblique aerial photograph of ice wedge polygonal ground, Arctic Coastal Plain, Alaska (photograph by Robert I. Lewellen).*

Pingos

Pingos are perennial frost mounds, consisting of a core, generally of massive ice produced principally by injection of water, and with a covering of soil and vegetation (fig. 15). Most pingos are domed to conical in shape, somewhat asymmetric, with a circular to oval base and a fissured top that may be cratered. The icy core of a pingo may take the form of a single, large lens of more or less pure ice or it may consist of a thick series of ice lenses interbedded with layers of frozen sediments, or a combination of these forms. The word *pingo* (Porsild, 1938) is a local Inuit word (Inuktitut language) for a hill, used in the Mackenzie Delta region of northwest Canada. The Russian term for pingo is *bulgunniakh*; it is derived from a Yakut word. Pingos occur in the continuous and the discontinuous permafrost zones, and the scars of former pingos (relict pingos), identified in mid-latitudes, are evidence of the previous extent of permafrost.

Two types of pingo are recognized (Müller, 1959): closed system and open system. Closed-system pingos (fig. 16) are typically found in areas of low-lying, poorly drained terrain in regions of continuous permafrost. In such regions, bodies of unfrozen ground (taliks) exist beneath the larger water bodies such as lakes and rivers. When a lake drains (Mackay, 1998), newly exposed sediments of the lake-bed cool, and permafrost begins to grow in this unfrozen soil (fig. 16). This permafrost aggradation creates high pore-water pressures within the unfrozen core, principally as a result of pore water being expelled from the freezing sediments. This high hydrostatic pressure is relieved by an upward movement of the ground surface. As the heaving continues over time, a mound that continues to grow is formed—the pingo (Mackay, 1979, 1998). The largest closed-system pingos known are about 50-m high, with side slopes as steep as 45°, and basal diameters of several hundred meters, but more typically they are smaller. Geothermal analyses suggest that pingos range in age from a few decades to many thousands of years.

The largest concentration of closed-system pingos is in the Mackenzie Delta region of northwestern Canada, where approximately 1,450 are known (Mackay, 1979). They are found also in northern Alaska, on the islands of the Canadian Arctic archipelago and, in much lesser numbers, elsewhere in

Figure 15.*—Pingo on the Tuktoyaktuk Peninsula, Mackenzie River delta, N.W.T., Canada (photograph by Harald Svensson, University of Copenhagen, Department of Geography).*

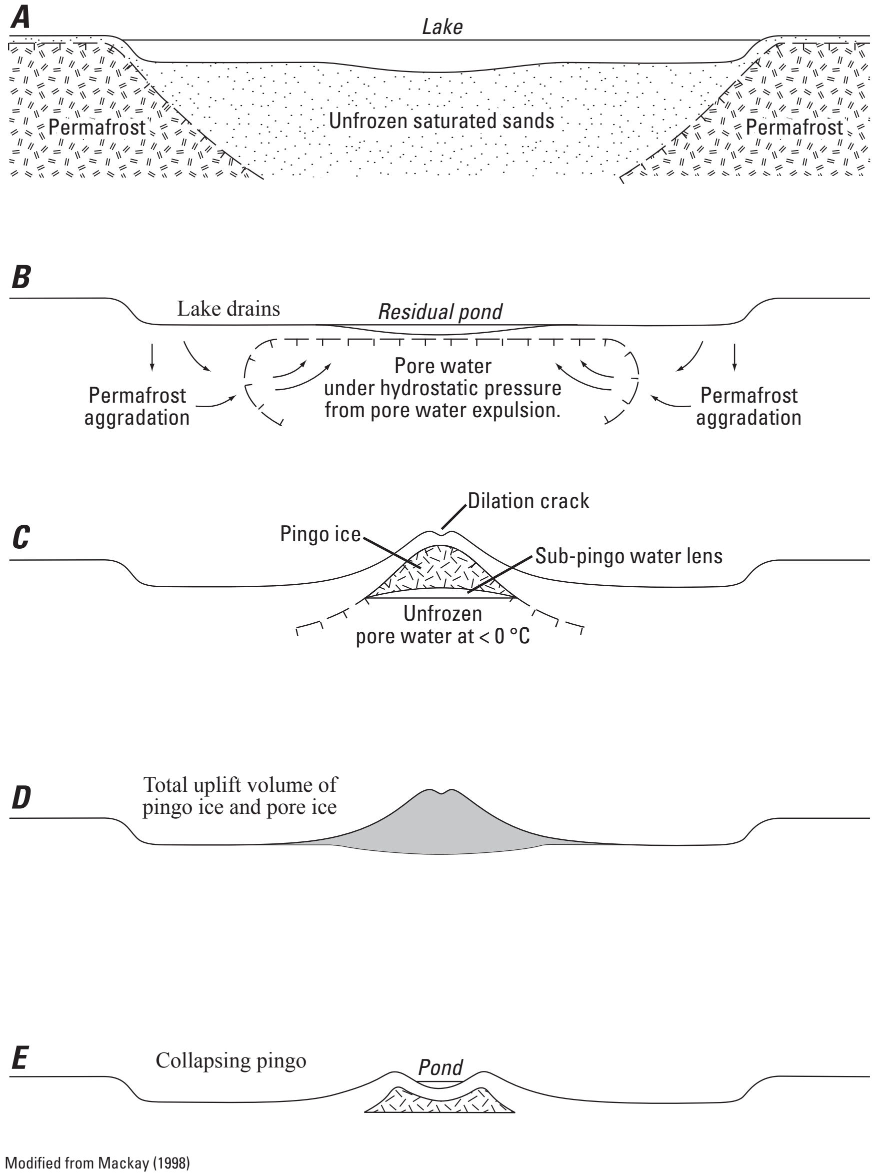


***Figure 16.**—Diagram illustrating the genesis and collapse of the closed-system pingos of the Tuktoyaktuk Peninsula area, Northwest Territories, Canada (from Mackay, 1998).*

northern Canada, including some in a submarine environment (Shearer and others, 1971). They occur along the north coastal regions of Russia and in the northern parts of western Siberia. Pingos also occur in certain coastal regions of Greenland and are frequent in Svalbard. A few small pingos are known from the Tibet Plateau.

Open-system pingos (fig. 17) tend to be smaller than closed-system pingos, even though examples from Svalbard attain heights in excess of 40 m (Svensson, 1971; Humlum and others, 2003). First described from Svalbard (Svensson, 1971; Liestøl, 1976), open-system pingos also occur in the ice-free areas of Greenland (Müller, 1959) and in the mountainous regions of northern Russia. In North America, they are most common in the Yukon-Tanana lowlands of central Alaska and in the central Yukon in areas that were unglaciated during the Wisconsinan glaciation. They are found in hilly to mountainous regions, where they occur in valleys and at the base of slopes, and primarily in areas of discontinuous permafrost with ground temperatures near 0°C.

Figure 17.*—Open-system pingo in upper Adventdalen, central Spitsbergen, Svalbard, Norway (photograph by Ole Humlum, University of Oslo, Department of Geography and The University Centre in Svalbard).*

Open-system pingos occur typically at the base of hill-slopes or on valley floors in areas of relatively thin permafrost. The ice core of an open-system pingo results from the injection of water from beneath or within the permafrost layer and under hydraulic pressure from the adjacent high ground (French, 1996; van Everdingen, 1998). The permafrost layer acts as an aquiclude and the sub-permafrost water is thus in an artesian situation. Seasonal frost mounds may be formed in a similar fashion in areas of deep seasonal frost.

Continued growth of a pingo typically leads to the rupture of the soil and vegetation cover across the summit. Should this expose the ice core, melting of the ice and thawing of the surrounding soil can lead to complete collapse of the pingo. Collapsed pingos are recognizable by their ringed shape; the resulting crater will often contain its own residual pond. Pingos near shorelines are also subject to erosion from retreating shoreline and wave action.

Palsas, Peat Plateaus, and Other Frost Mounds

Frost mounds, a few meters in height and from 5 to 20 m in diameter, are smaller than pingos, and the frozen core is generally formed of ice lenses in frozen peat and soil. Frost mounds include palsas (fig. 18), peat plateaus, and frost blisters. The term *palsa* is originally a Sami-Finnish word (Finno-Ugric language) for a dry mound or ridge situated in a bog, but it is now internationally used as a geomorphological term. Palsas are typical of the discontinuous zone of permafrost and consist of a peaty mound with a core of ice lenses and frozen peat or soil formed in a wetland environment (Friedman and others, 1971). The growth of the mound is due to the development of the ice layers. A peat plateau is similar to a palsa but more extensive in area. The frozen cores of palsas and peat plateaus are protected from thawing during the summer by an insulating dry surface layer of peat (Åhman, 1977). Palsas have a domed form, 5 to 25 m across, and they often grow together, forming an esker- or plateau-like complex of peat mounds; in Scandinavia they can be as long as 500 m (Svensson, 1970). Their maximum height is 6 or 7 m. A frost blister is a seasonal frost mound with a core of ice that forms under conditions of high hydraulic pressure during the freezing of the active layer.

Figure 18.—*A 5.5 m-high palsa in Varangerfjord area, northern Norway (photograph by Harald Svensson, University of Copenhagen, Department of Geography).*

Rock Glaciers

Two main types of rock glacier are recognized: *talus-derived rock glaciers*, which occur below talus slopes, and *glacier-derived rock glaciers*, which are found below mountain glaciers or below areas that such alpine glaciers used to occupy (fig. 19). The ice component of a rock glacier is normally at least 60 percent of the rock glacier's volume and thereby exceeds the natural void volume of the rock fraction, thus allowing deformation of the ice-rich matrix and providing a mechanism for the rock glacier's movement.

Rock glaciers are characteristic large-scale flow features of frozen material in high-relief permafrost regions (Haeberli, 1985; Calkin and others, 1987; Barsch, 1996; Humlum, 1998; Humlum and others, 2007; Haeberli and Gruber, 2008). They are located at the foot of rock faces with an abundant supply of talus and, when active, typically take the form of 20- to 100-m-thick tongue- or lobe-shaped bodies with cascading frontal slopes standing at the angle of repose. Rock glaciers may be as much as several kilometers in length, but the typical length is 200 to 800 m measured parallel to the direction of flow (Barsch, 1996). The surface of rock glaciers is covered by coarse (0.2 to 5 m) rock fragments, and its arcuate transverse furrow-and-ridge topography ranges in height from 1 to 5 m. Active rock glaciers typically flow downslope at the rate of 0.1 to 1 m a^{-1} (see, for example, Barsch, 1996); that is, they exhibit a notably slower flow rate than normal glaciers. The slow movement leads to the development of arcuate furrows and ridges on the surface, typically aligned across the main direction of movement. When a tongue-shaped rock glacier reaches a valley floor, it may spread out into a spatulate form.

Active rock glaciers are often construed as characteristic of continental environments, although several observers have described rock glaciers from maritime regions (Svensson, 1989; Sollid and Sørbel, 1992; Humlum, 2000). Corte (1987, 1998) presented a taxonomic classification of active rock glaciers at several locations in the Andes and photographed a rock glacier in Argentina (figs. 20, 21).

An active rock glacier is more or less in equilibrium with its present environment. It may become inactive for one or more reasons. Climate change may reduce the supply of snow and rock fragments to the head of the rock glacier. The surface mantle of ice-free rocks (the active layer of the rock glacier) may become sufficiently thick and massive to prevent any movement. Finally, the ice-rich layer may become too thin to support further deformation and transport of the mantle. The complete loss of the ice-core matrix will result in a relict rock glacier, recognized by the presence of collapse features on the surface.

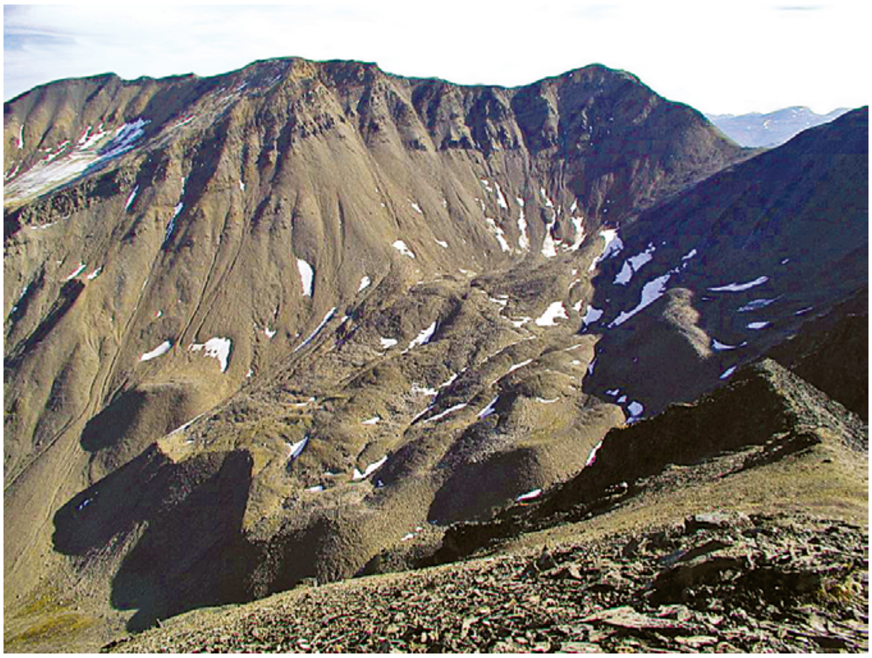


Figure 19.*—Rock glacier in Atigun Pass area, northern Brooks Range, Alaska (photograph by Atsushi Ikeda, University of Tsukuba, Japan).*

Figure 20.*—Morenas Coloradas rock glacier in Argentina (photograph by D. Trombotto; from Romanovsky, Gruber, and others, 2007).*

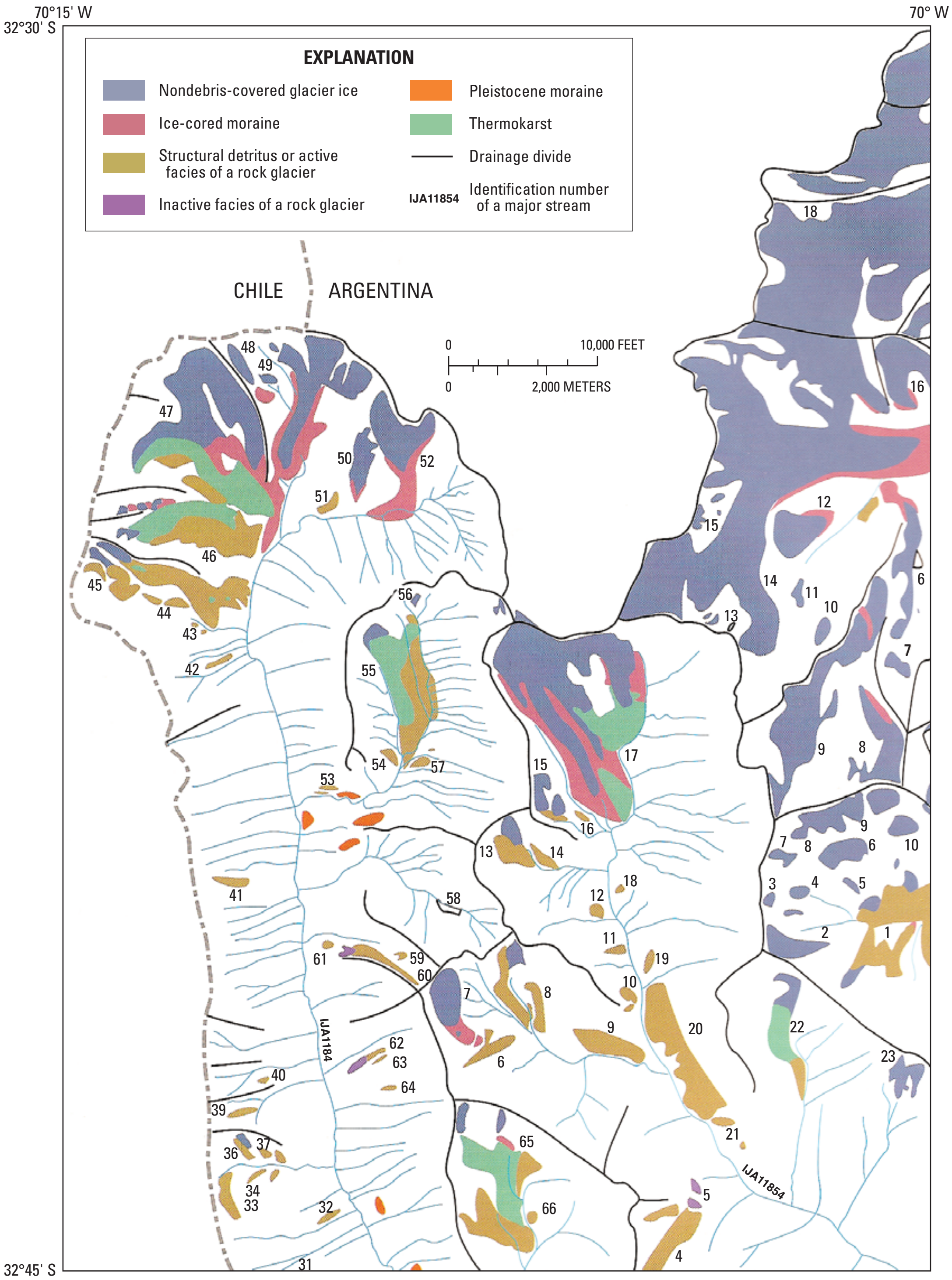


Figure 21.—*Distribution of rock glaciers, glaciers, ice-cored moraines, and thermokarst features, Mendoza basin, western Cerro Aconcaqua, Argentina (Corte, 1998; fig. 18, p. I139–I141). Numbers refer to individual glaciers and rock glaciers assigned by Arturo Corte.*

Thermokarst Processes and Landscapes

The so-called karst phenomenon, the morphology developed by the melting of ground ice, is termed *thermokarst* (Ermolaev, 1932), by analogy with the phenomenon that produces dissolutional landforms in limestone areas. The thawing of ice-rich permafrost beneath the active layer may strongly alter the surface morphology, resulting in subsidence and cave-in formations of the ground surface. When ice-rich permafrost thaws, the ground surface consolidates. If thawing proceeds relatively slowly and the texture of the soil allows the meltwater to drain away, the ground surface will be lowered by an amount related to the excess ice content—a process referred to as "thaw subsidence." If, however, thawing occurs more rapidly than meltwater can be expelled from the thaw zone, a quick (ground failure) condition can develop, with the water carrying part of the overburden load. The result is a complete loss of strength within the soil mass, which then behaves as a viscous liquid.

Thermokarst terrain forms wherever ice-rich permafrost thaws and the ground surface subsides into the resulting voids. The amount of ground ice initially present, the form in which the ice was present, and the rate of thaw are the conditions that determine which form of thermokarst develops. Thus melting of segregated ice, pingo ice, wedge ice, and massive ice may each result in different landforms. On level terrain, natural thermokarst typically begins as one or more small depressions, commonly containing meltwater, and often developing at the intersections of ice-wedge polygons (Jorgenson and others, 2006) (fig. 22). These small thermokarst ponds enlarge by thermal erosion thawing of the surrounding banks; if wedge ice or massive ice is exposed to the air or water, this enlargement can be very rapid. Beaded streams (fig. 23) or pitted drainage forms when ice wedges melt at their intersection under the influence of stream runoff. Adjacent thermokarst ponds may coalesce, forming thaw lakes. Thermokarst depressions, such as thaw lakes (figs. 24, 25) and alas areas ("A thermokarst depression with steep sides and a flat grass-covered floor"; Neuendorf and others 2005, p. 13) (fig. 26), and some collapse features in Arctic areas are so distinct in form and contour that they are considered as good indicators of the presence and characteristics of the surrounding permafrost terrains.

Figure 22.*—Initiation of thermokarst by melting of ice at intersections of ice wedges, northern Alaska (photograph by M.T. Jorgenson, ABR, Inc., Fairbanks, Alaska).*

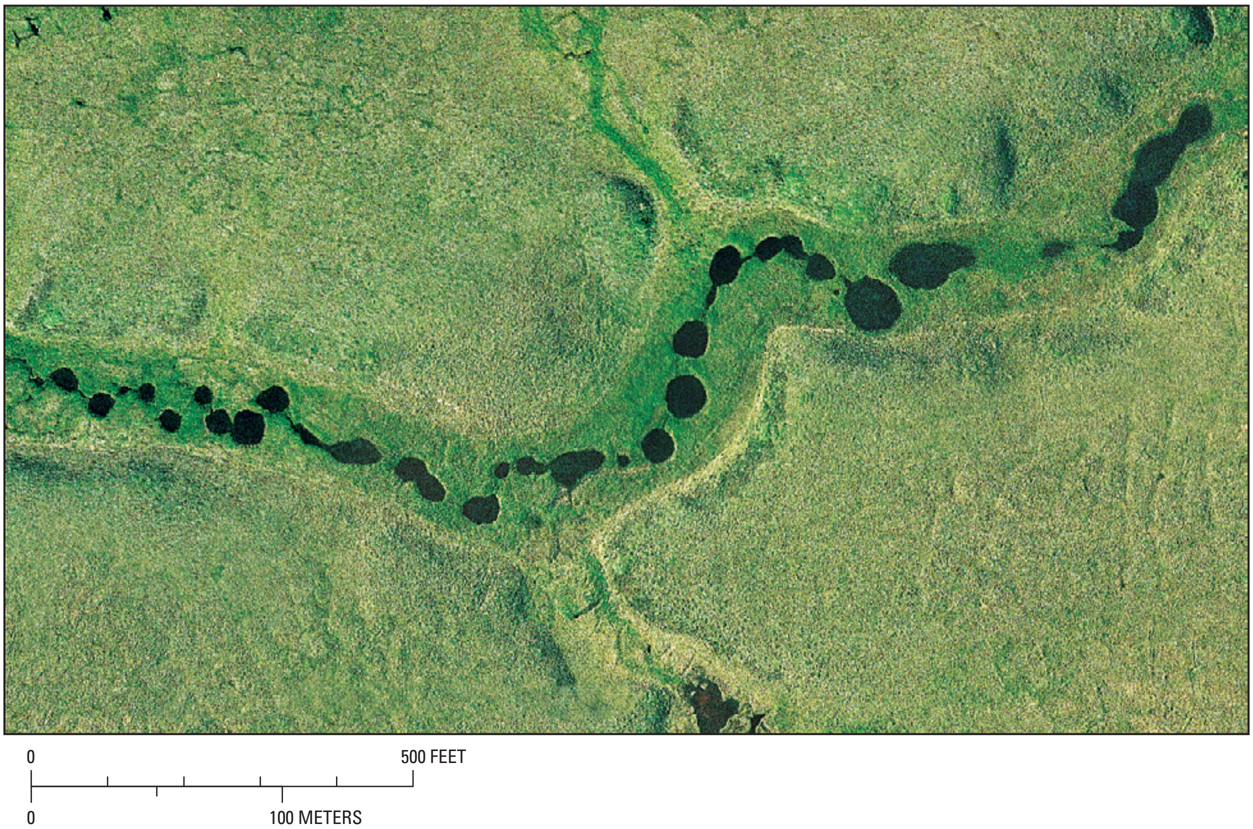


Figure 23.—*Vertical aerial photograph of beaded stream channels resulting from melting of ice wedges and thawing of surrounding ice-rich permafrost. The photograph was taken in summer 2004 south of the town of Nuiqsut, northern Alaska, by M.T. Jorgenson, ABR, Inc., Fairbanks, Alaska.*

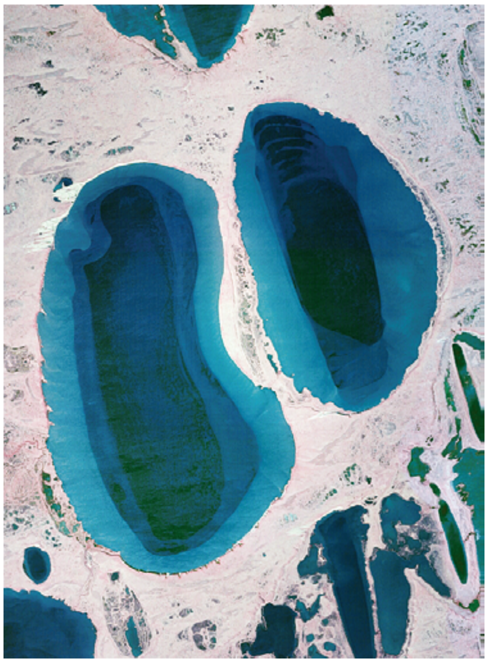


Figure 24.—*Thaw-lake basins on the Arctic Alaska Coastal Plain. The central deep basins result from the thaw of ice-rich sediments with a resulting underlying thaw bulb (see fig. 2) (photograph by Ben Jones, U.S. Geological Survey).*

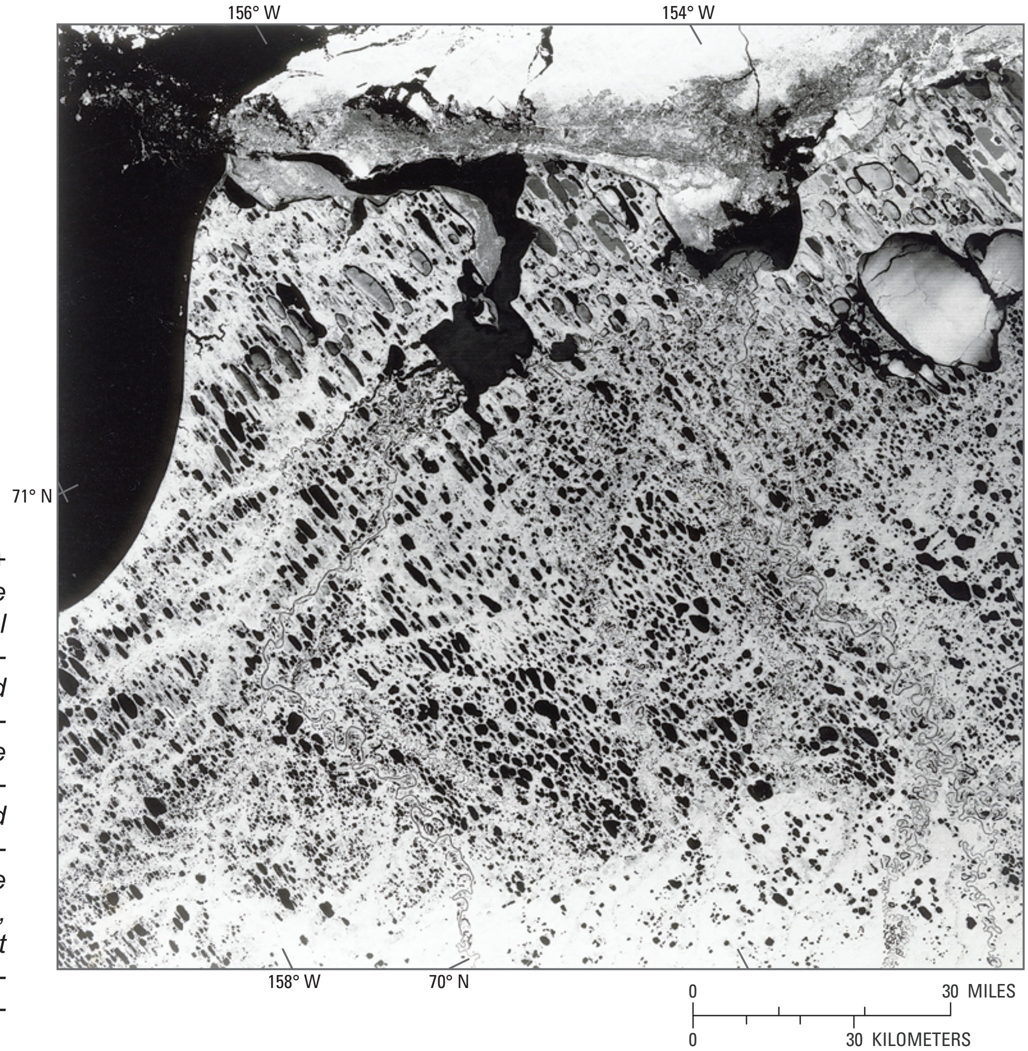


Figure 25.*—Landsat 7 ETM+ false-color composite image of the Alaskan Arctic coastal plain showing a predominance of elongated oriented lakes and other forms of thermokarst lakes. The lakes are oriented approximately north-south; the prevailing wind direction is from the northeast. Landsat 7 ETM+ image (L71079010-01020000815, path 79, row 10; 15 August 2000) from the U.S. Geological Survey EROS Data Center, Sioux Falls, S. Dak.*

Figure 26.*—Alas valley resulting from thawing of ice-rich permafrost, central Siberia, Russia (photograph by Vladimir Romanovsky, University of Alaska Fairbanks).*

Thaw Lakes

Thermokarst processes are responsible for thaw lakes, the water-filled depressions in many coastal or alluvial plains in permafrost areas. Thaw lakes depend on the ice content of the ground, and so they develop optimally in areas of ice-rich, fine-grained silty and sandy soils (Sellmann and others, 1975; Jorgenson and Shur, 2007). Although the dimensions of thaw lakes may differ in the same area, their shorelines are regularly circular to oval except where thermo-erosive activity enlarges the lake to intersect with adjoining lakes.

Thaw lakes characteristically drain, often suddenly, the outflow following lake tapping or truncation by coastal retreat. Thermal erosion along ice wedges frequently leads to lake tapping.

Groups of thaw lakes with pronounced parallelism of their long axes ("oriented lakes") are known from several areas in northern Alaska, Canada, and Russia. The oriented lakes of northern Alaska have been described and classified, and hypotheses have been presented to explain their origin and orientation (Carson and Hussey, 1962; Price, 1968; Sellmann and others, 1975; French, 1996; Hinkel and others, 2005). Figure 25 illustrates these thaw lakes on part of the northern coastal region of Alaska, where the north-northwest to south-southeast orientation of the elliptical lakes is very pronounced. Such a preferred orientation prompts discussion of the causal geomorphic agent. The influence of wind has been proposed, but the predominant wind direction is at right angles to the long axis of the lakes. The possible effect of insolation, the structure of the bedrock, and thaw slumping have also been evaluated. A convincing hypothesis that explains the preferential orientation of the thaw lakes, however, has not yet been accepted by all scientists. Recent observations from the northern area of western Siberia show that the abundance and area of thaw lakes are declining (Smith and others, 2005).

Alasses

Some terraces along Siberian rivers are covered with oval thermokarst depressions that may reach 10 m to as much as 15 km in length. Such a depression is termed an "alas" (fig. 26), a Yakut word (Turkic language) in origin. Viewed from the air, alasses stand out as open, grass-covered, closely spaced, and flat-bottomed basins in the taiga, often containing lakes or swampy areas, which have developed from former lakes. The alas landform develops in the course of a broad spectrum of small-scale morphological processes and may also involve the formation of pingos (bulgunniakhs) in the successively drained or sediment-filled alas lakes (Solovyev, 1973). A climatically induced increase in the depth of the thaw favors the development of alasses but, as with other thermokarst features in permafrost, the alas may begin forming without any climatic change. Local disturbances of the vegetation cover may be sufficient to start the process, the most effective agents being forest fires and floods. Some of the larger systems of alasses in central Yakutia may have begun to form during climatically favorable periods of the late Pleistocene or Holocene Epochs.

Slope Failures and Solifluction

Freezing and thawing of slope deposits in ice-rich permafrost and related slope failures produce specific landforms: detachments, thaw slumps, and solifluction (figs. 27, 28, 29).

Detachment Failures

Detachments occur when a thawed active layer and vegetation mat separate from the underlying permafrost table. The causes of such failures may include an unusually warm summer with deeper than normal thawing of the ground, wildfire, or unusually heavy rain. The resulting landform is typically spatulate, with the detachment area on the upper part of the slope and a lobe of

Figure 27.—*Large retrogressive thaw slump, northwest Alaska, triggered by lateral erosion of the Selawik River (photograph by Kenji Yoshikawa, University of Alaska Fairbanks).*

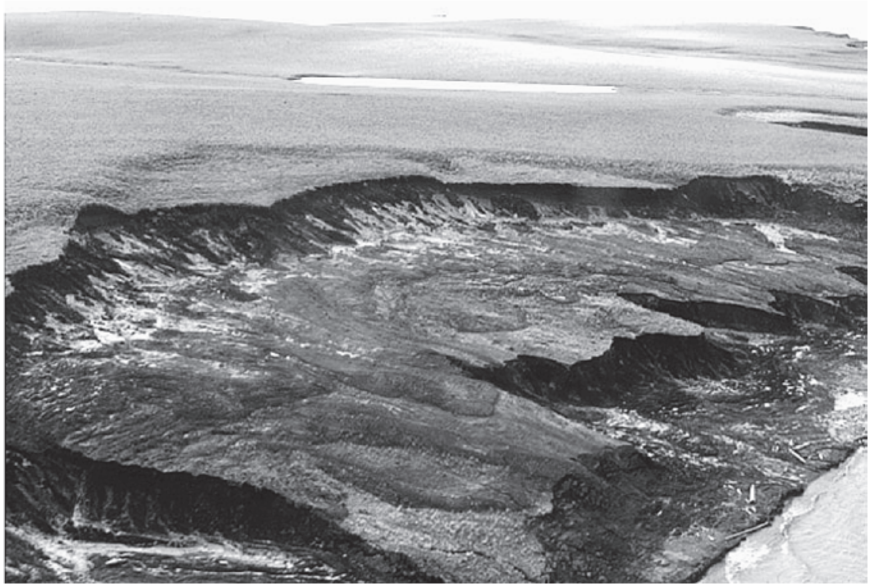


Figure 28.*—Thawing permafrost, western Canadian Arctic (photograph by Hugh French).*

Figure 29.*—Solifluction lobes, north of Griegdalen, Svalbard, Norway (photograph by Ole Humlum, University of Oslo, Department of Geography and The University Centre in Svalbard).*

debris downslope. Detachment failures often occur in groups on south-facing slopes with angles of 12° to 14° at their source. They can develop on slopes with angles as low as 6°, and they can traverse slopes as low as 1° or 2°. Should this initial failure expose massive ice or ice-rich soil, a thaw slump is likely to develop subsequently.

Thaw Slumps

These slope failures are deeper and become larger where ice-rich soil or massive ground ice is exposed at the ground surface. A thaw slump forms a bowl in a slope with a steep headwall in which the massive ice or ice-rich soil is exposed. During the summer thaw, the floor of the bowl forms a debris flow of soupy mud flowing away from the ice face. As the ice face melts and retreats, the soil of the active layer slides down the ice face, mixing with the accumulated meltwater and contributing to the debris flow. Large, compound thaw slumps can be tens of meters in depth and as much as several kilometers in width. Such slumps will expand retrogressively as long as massive ice or icy soil is exposed in the headwall.

Solifluction

Solifluction is the slow downslope flow of saturated earth materials. During seasonal or daily freezing, the formation of ice lenses and needle ice causes soil surfaces to expand and heave upwards, the soil particles moving upwards perpendicular to the surface. Subsequent thawing of the ground allows the soil to resettle in a more nearly vertical direction. Downslope movement occurs because the heave and resettlement involve this ratchet-like motion. The overall effect of these processes is thus a small diagonal or downhill displacement (Matsuoka, 2001).

Solifluction leads to the formation of lobate-shaped debris masses called solifluction lobes (fig. 29). They are typically 10 to 100 m in length, 5 to 100 m in width, and display steep frontal margins or risers up to 1.5 m in height. Because frost-sorting processes operate simultaneously on active solifluction lobes, a concentration of large clasts can typically be seen around their outer margins. Such solifluction lobes are “stone-banked.” Lobes that lack marginal clasts are “turf-banked.” Areas of laterally extensive solifluction lobes are referred to as “solifluction sheets.”

Other Periglacial Forms and Processes

In addition to ground freezing and thawing processes, which are important because of their magnitude or frequency, and their widespread occurrence (Walker, 2005; French and Thorn, 2006), other phenomena of special interest to periglacial research include seasonal snow cover as an important source of moisture and river-, wind-, and coastal-related processes. Rivers in periglacial regions usually exhibit specific temporal patterns of discharge. Most of the flow tends to occur during the weeks of snowmelt (the nival flood). The concentration of discharge in a short period, in combination with a heavy sediment load, results in poorly developed, shallow, braided channels. At the ground surface in many periglacial areas, wind speeds are characteristically high, partly because tall vegetation is absent. Such winds often set in motion large quantities of loose sediment, including snow and soil. Subsequent depositional features include sand dunes, loess accumulations, and sand sheets. Erosional processes include coastal and river bank retreat and wind-polishing of boulders and of rock outcrops.

Patterned Ground

Intense seasonal frost action, in the form of repeated freezing and thawing, produces small-scale patterned ground; circles, stripes, or polygons develop on the ground surface. Patterns having coarse sediments, such as stones, bordering them are referred to as "sorted," and patterns without accompanying stones are "nonsorted." By 1956, Washburn (1956) was able to list 19 hypotheses for how patterned ground develops. Repeated freezing and thawing sort granular sediments, forming circles (fig. 30), stone nets, and polygons. The coarse cobbles and boulders form the outside of the ring, and the finer sediments form the center (Hallet and Prestrud, 1986; Werner and Hallet, 1993; Hallett and others, 2004). At a fine scale, local topography, soil, hydrology, snow thickness, microtopography, and soil moisture control the distribution

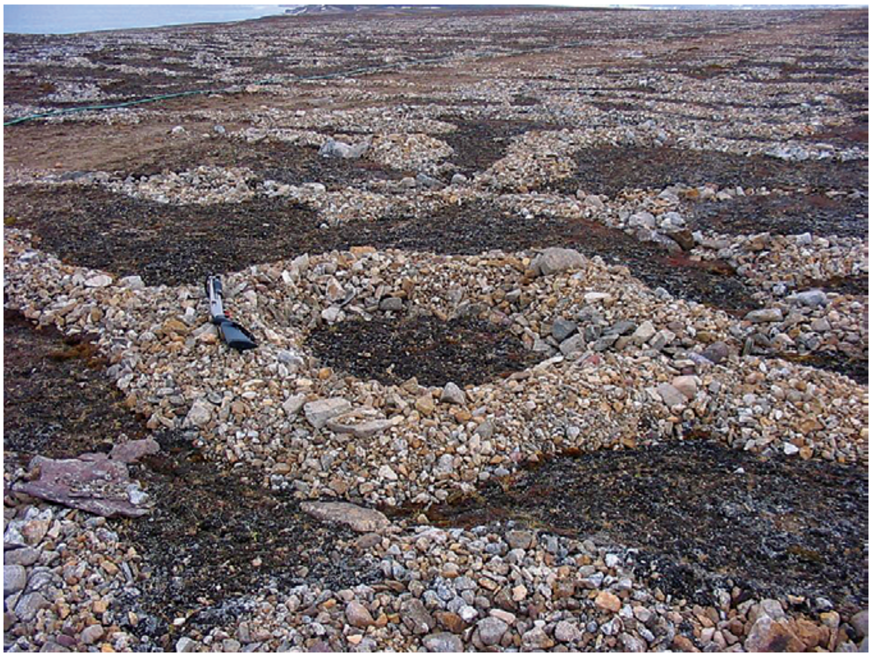


Figure 30.*—Patterned ground as illustrated by sorted circles (see rifle for comparison of size), Kongsfjorden-Brøggerhalvøya area, Svalbard, Norway (photograph by Grzegorz Rachlewicz,Uniwersytet im. Adama Mickiewicza, Poznan, Poland).*

and formation of patterned ground (Washburn, 1980; Matthews and others, 1998; French, 2007). Patterned ground in the form of sorted circles is widespread on level ground in permafrost regions. On slopes, the circles become elongated downslope, forming stripes. The dimensions of patterns vary greatly. The diameter of stone circles and polygons ranges from 0.5 to 50 m, and stripes may be 1 to 10 m in width and 100 m or more in length. Circles and polygons form on horizontal surfaces having maximum gradients of 5°, whereas stripes develop on slopes with gradients of 5° to 30°. Active, large-scale (greater than 1 m) patterned ground typically signals the presence of permafrost below, but the presence of permafrost is not required in order for small-scale patterned ground to form.

Non-sorted circles, often in the form of hummocks, or vegetated mounds, are found in a variety of environments (Washburn, 1980). Large stone stripes occur on slopes, where the circular or polygonal pattern becomes elongated. The stripes are generally oriented down-slope, as opposed to across the slope. Small stripes "resembling the pattern formed by a garden rake" are widespread in many periglacial regions as well as outside permafrost regions.

Detailed studies of the dynamics of well-developed sorted circles led Hallet and Prestrud (1986) to conclude that convection moves sediment in the fine-grained center in sorted circles and "subduction" moves it where fine and coarse grains are in contact. A numerical model for self-organization of sorted patterned ground was proposed by Kessler and Werner (2003), drawing on examples from field studies in Kvadehuksletta, Svalbard, Norway.

Nivation Hollows

Periglacial processes enhanced by the presence of snow create nivation hollows as common phenomena (fig. 31). These features can develop under snow patches in just a few seasons. Nivation hollows typically occur on a slope that faces downwind in relation to wind direction in winter (Christiansen, 1998). The hollows require a snow patch that persists or reforms in the same area year after year and a slope that allows for erosional transport of material out of the developing depression. Around the edges of the snow patch, physical weathering and frost heaving begin to separate particles and thus prepare them

Figure 31.*—Nivation hollow, Disko Island, Greenland (photograph by Ole Humlum, University of Oslo, Department of Geography and The University Centre in Svalbard).*

for erosion by flowing water. As summer progresses, the patch of snow reduces in size and the excavation of material continues inward. In the following year, the boundary of the snow accumulation and the depression will once again coincide, and sliding and frost weathering will erode the hollow's edge. The edge of the hollow is further eroded because the microslope creates localized instabilities that focus the entrainment potential of flowing water.

Icings (Naledi, Aufeis)

In permafrost areas, vast surficial sheets of ice develop during the fall and winter by freezing of river or ground water that has flowed across cold ground or snow surfaces. The sheets build up large accumulations of ice in parallel layers (fig. 32). Warm glaciers discharge water during the winter and may also build up thick accumulations of ice beyond their termini. When such ice accumulations are observed in summer, they may be incorrectly interpreted as perennial snow fields because of their bright appearance. The Russian word *naledi* has long been used for this special type of ice accumulation; in German the term *aufeis* is equivalent. The corresponding Inuit word in Greenland is *sersinaq* (Humlum and Svensson, 1982). The common English term for this phenomenon is *icing*.

Figure 32.*—The face of an icing (5-m high) along Echooka River, northern Alaska, 11 July 1972 (Sloan and others, 1976).*

Using observations in West Spitsbergen, Svalbard, Norway, Åkerman (1980) published a classification of naledi based on the origin of the water involved: (1) spring naledi (water from ordinary ground water or from thermal springs), (2) pingo naledi, (3) stream or river naledi, and (4) glacier naledi. Most naledi melt, or they become strongly dissected during the summer, but larger ice accumulations persist from one season to the next and may influence the local climate and even cause damage to the adjacent vegetation. Very large naledi occur in intermontane basins of northeastern Siberia. In the Momskaya depression the largest naledi are about 100 km^2 in extent and reach a thickness of 5 to 6 m. Using Landsat MSS images, Williams (1986) described icings in northern Alaska (fig. 33). Two prominent icings are described by R.M. Krimmel on a 29 August 1978 Landsat 3 MSS image (30177-20435; Path 76, Row 11) in the eastern part of the Brooks Range (see fig. 438, p. K472, Glaciers of Alaska [1386–K] of this series). Other large icings along the Trans-Alaska Pipeline route in Alaska and in eastern Greenland are described by Sloan and others (1976) and by Humlum and Svensson (1982), respectively.

Thermoerosional Niche

Where ice-rich soils are exposed in cliffs along the Arctic coast and along Arctic rivers, wave erosion can lead to undercutting of the cliff face, forming a thermoerosional niche (fig. 34) that can be many meters deep (Jorgenson, 2009). Under the tensional strain, the overhanging frozen ground of the cliff eventually collapses. If the collapse occurs along the line of an ice wedge, a large block of frozen soil falls onto the beach or riverbank. The fallen blocks of frozen ground initially protect the cliff face from further undercutting, but thermal abrasion thaws and erodes this material until a fresh face is exposed. This process is sensitive to floods on rivers and to storm surges and high wave action in the coastal environment. Late summer storms have the most damaging effect, because an ice-foot or snow bank generally protects the foot of the cliff earlier in the thaw season. Longer periods of ice-free conditions and more frequent summer storms increase the likelihood of coastal erosion.

Windblown Deposits (Sand Dunes, Sand Sheets, and Loess Deposits)

Wind has conspicuous effects on the open landscape and sparse vegetation in cold climates. The Arctic is characterized by wind-polished rocks, wind-oriented lakes, sand dunes, and loess deposits that owe their formation to eolian processes controlled by frost and snow cover.

Dunes, sand sheets, and loess deposits may form along coasts and river plains. The main wind action along rivers typically takes place during the autumn, when braided river plains tend to dry up because of low temperatures and decreasing volume of discharge before the winter snow cover has become established. Dust storms then may result in meter-thick accumulations of eolian deposits along such braided river plains, reflecting the prevailing wind direction.

Where a steep rock cliff is exposed to strong winds, fine-grained weathering products may be blown up the headwall to be deposited in sheets or dunes near the summit. Such periglacial eolian deposits are known from Scotland and from the Færoe Islands (Ballantyne and Whittington, 1987; Humlum and Christiansen, 1998).

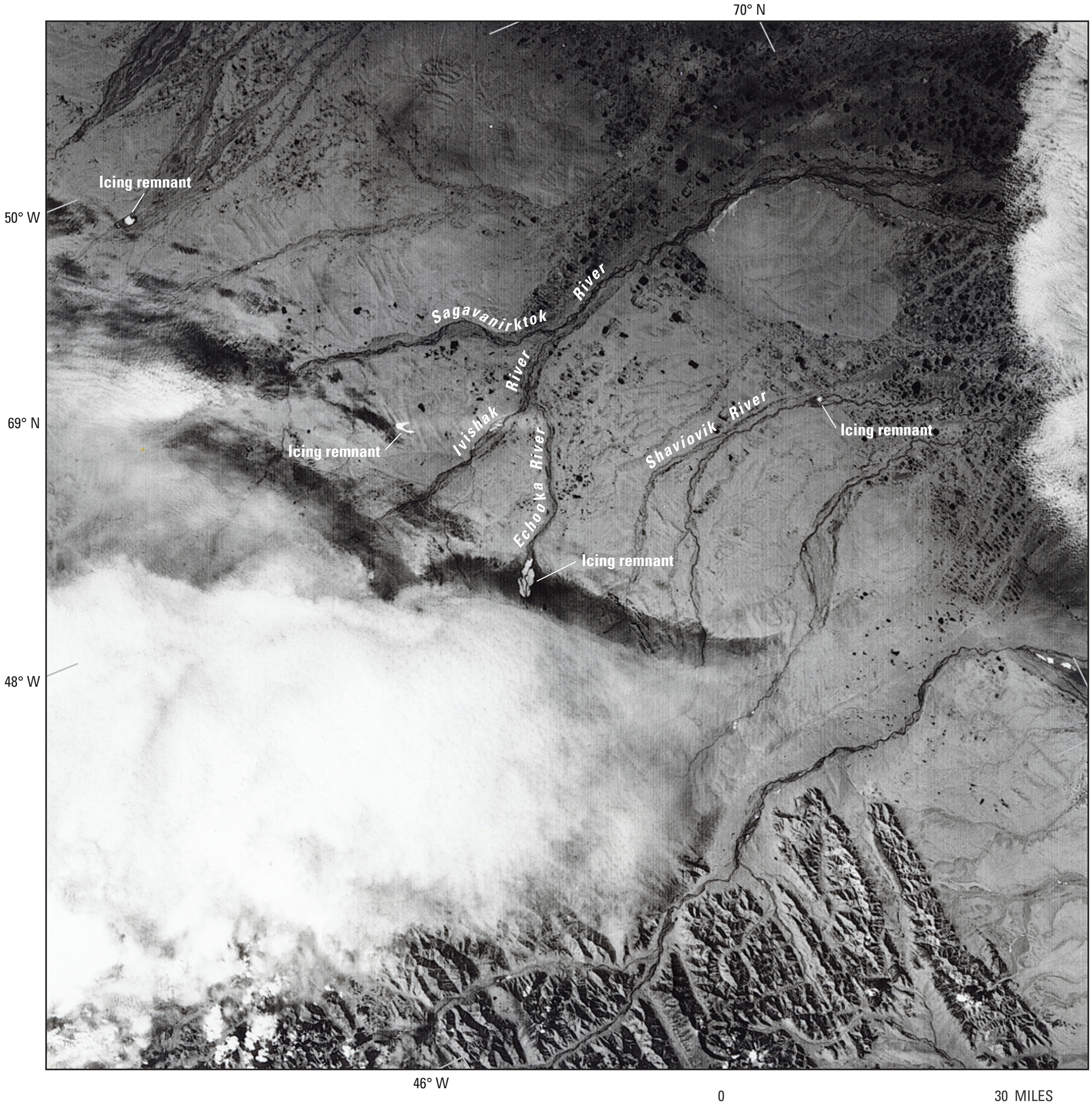


__Figure 33.__—Landsat 1 MSS image of icings in northeastern Alaska on 4 August 1973 (fig. 9-24 in Williams, 1986). Landsat 1 MSS image (1377-2112, band 6) from the U.S. Geological Survey EROS Data Center, Sioux Falls, S. Dak.

Figure 34.—*Coastal erosion initiated by undercutting of the bank (niche) that subsequently results in block collapse. Several ice wedges are exposed along the Beaufort Sea at Pitt Point, coast of northern Alaska (photograph by M.T. Jorgenson, ABR, Inc., Fairbanks, Alaska).*

In very arid areas with permafrost, extreme winter temperatures can result in ground cracking by thermal contraction, as in the formation of ice wedges. Windblown fine sand and silt can be deposited in these cracks, leading to the growth of sand wedges and polygonal patterned ground. Excellent examples of sand wedges associated with ice wedges and composite wedges of sand and ice are known from the McMurdo Dry Valleys region of Antarctica (Black and Berg, 1966; see also references in Bockheim, 1995).

Wind action can shape and polish clasts (for example, boulders and cobbles), forming ventifacts. Blowing silt, sand, and even snow can polish outcrops of bedrock. Finally, the winnowing of fine particles from the ground surface results in surfaces covered with lag gravels.

Fossil or Relict Periglacial Features

From a palæoclimatic point of view, it is important to identify fossil periglacial features in the landscape. Attempts at reconstructing past environments are based primarily on studies of certain fossil periglacial landforms and sedimentary structures, such as casts of ice wedges and sand wedges, thermal cracks, fragipan horizons (hard, low permeability subsurface layer), the systematic occurrence of cryoturbation (soils or unconsolidated sediments displaced by frost action) or solifluction, evidence of past strong ice segregation in sediments, and fossil rock glaciers. Climatic changes resulting in the areal disappearance of ice-rich permafrost (for example, beyond the margins of the previously extensive Quaternary Period ice sheets and ice caps), tend to blur or confuse the evidence of even the most distinct periglacial features. Traces of their morphology and relict surface and subsurface structures, however, may be detected in many areas (fig. 35). These traces, often quite subtle, substantiate the fact that they were originally periglacial features that are now in a fossil or relic stage (Lundqvist, 1962; Svensson, 2005). Such fossil forms, especially when dated, are used as indicators of a previous, much colder climate of the area (Péwé, 1973; Vandenberghe, 2001).

Figure 35.—*Relict pattern of ice-wedge polygons on the Laholm Plain, near the west coast of Sweden (photograph by Harald Svensson, University of Copenhagen, Department of Geography).*

Seasonally Frozen Ground

Only sparse historical observational measurements of the depths of seasonal freeze and thaw are available. Regions of seasonally frozen ground occupy approximately 24 percent of all permafrost regions and approximately 50 percent of the exposed land surface of the Northern Hemisphere (Zhang and others, 2003) (seasonally frozen ground, fig. 36). Despite this vast area that seasonally frozen ground occupies, particularly outside the region of permafrost, relatively little research has been undertaken on the importance of seasonally frozen ground as a factor in the major processes typical of seasonally cold-climate environments. One reason for this neglect continues to be the difficulty of measuring and estimating the areal extent of surface frost that occurs under seasonal snowpacks. Seasonal freezing and thawing of the ground, however, clearly have important effects on hydrological, ecological, and geomorphic processes by reducing infiltration and increasing water runoff. Soil temperature and freeze/thaw status also influence agricultural productivity, native plant growth, carbon exchange between the land surface and the atmosphere, surface and subsurface hydrology, and weather and climate through the surface-energy balance.

In middle and high latitudes, the importance of the timing, thickness, and duration of seasonal snow cover on the extent of seasonally frozen ground is considerable. For temperate regions the overall effect of snow cover (unless it is thin) is to insulate soil and keep soil water from freezing. During the early winter, the area of frozen soil in regions with little snow cover dominates the total area of seasonally frozen soil. As winter progresses and the snow-covered area expands, frozen soil under snow cover becomes dominant. By the end of winter, soil frozen because of the insulating effect of snow may persist into the early summer until the snow disappears.

Even a thin frozen layer can prevent the exchange of moisture between the atmosphere and soils. Frozen soil containing a large volume of ice has a large apparent heat capacity, consuming or releasing about 80 times the amount of

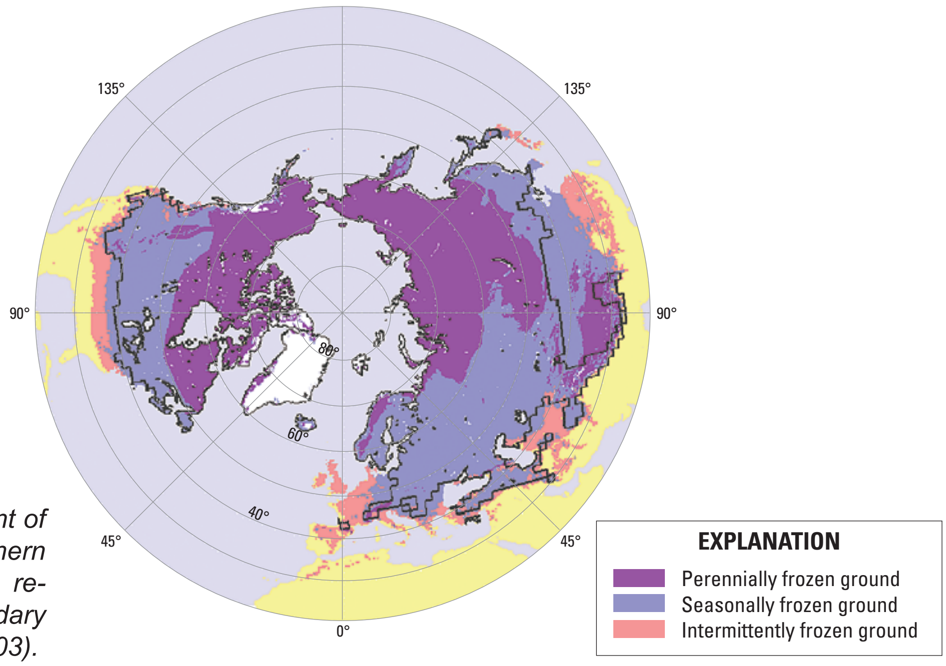


Figure 36.*—Map showing the extent of seasonally frozen soils of the Northern Hemisphere; both permafrost and regions south of the permafrost boundary are included (Zhang and others, 2003).*

energy in freezing or thawing than water consumes or releases during heating and cooling. Additionally, the processes of freezing and thawing can affect the decomposition of organic substances in the soil as well as the migratory patterns and physiology of biota in the soil. Although greenhouse-gas exchange between the atmosphere and the land surface may be minimal when soil is frozen, the period following spring thaw shows a marked release of greenhouse gases, known as a "respiratory burst."

Seasonal freezing and thawing processes in cold regions play an exceedingly important role in diversity and productivity of the ecosystem and in the polar and subpolar hydrological cycle in general. Furthermore, long-term changes in seasonal depths of freeze and thaw are important indicators of climate change.

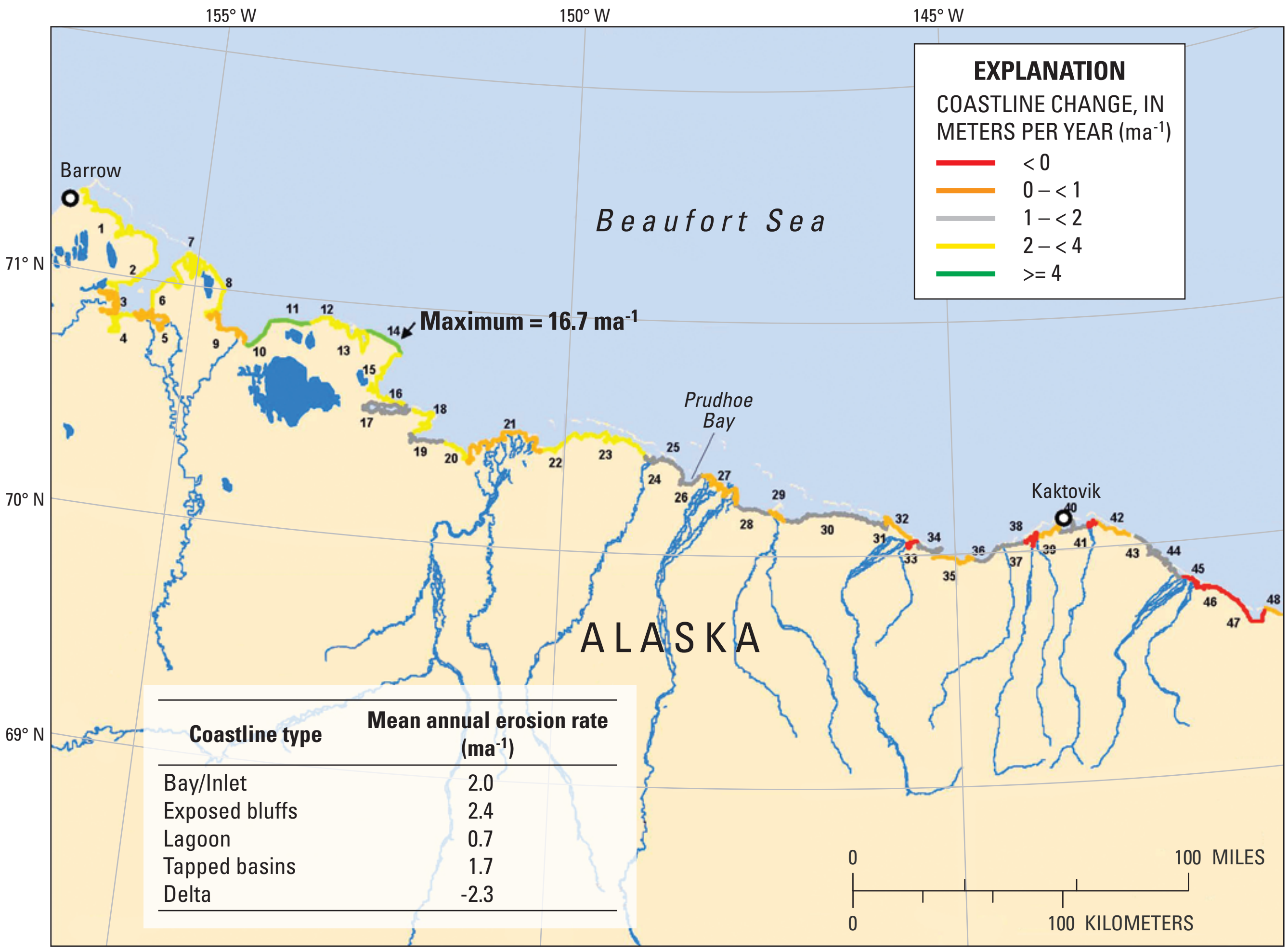
In high-altitude areas of middle and low latitudes, diurnal freezing and thawing of the ground have more far reaching effects than seasonal freezing. Patterned ground can develop in these areas, but on a much smaller scale than in polar regions.

Remote Sensing of Permafrost and Periglacial Landforms

Airborne or satellite remote-sensing platforms cannot directly detect or measure properties of permafrost terrain, and especially they cannot detect whether permafrost is present or absent (Brown and others, 2007). Many surface features of permafrost terrains and periglacial landforms are observable from conventional aerial photography and from high-resolution satellite imagery. Surface indicators of permafrost terrains that are discernible visually and thus by airborne or satellite remote sensing, or both (assuming sufficient spectral discrimination and picture-element (pixel) resolution), include pingos, thaw lakes and basins, ice-wedge polygons (fig. 14), beaded drainage (fig. 23), palsa fields (fig. 18), slope failures (figs. 27, 28), and rock glaciers (figs. 19, 21). Vegetation and soil moisture can also indicate differences in the subsurface permafrost conditions. The former extent of permafrost may be inferred from fossil features that are observable from the air in regions where permafrost no longer exists (fig. 35).

Remote sensing and geographic information system (GIS) technologies provide suitable observational and mapping methods for identifying many types of periglacial features, for analysis of their areal distribution, and, in some cases, for calculating their rate of formation and subsequent change. Actively developing periglacial phenomena as well as periglacial features in various stages of degradation commonly are distinctly outlined and regularly grouped, giving a characteristic pattern to their ground appearance. This distinctive appearance, together with the fact that most periglacial areas have no forest or high-standing vegetation cover (except for boreal forest and taiga), account for the relatively straightforward identification of periglacial morphology by surface features, provided the observer has adequate experience in the field. Photographic records, especially medium-scale (medium-altitude) vertical aerial photography, have been the most common imaging method. Although the spatial resolution of small-scale, high-altitude aerial photographs and the pixel resolution of readily available satellite images may limit or preclude the detection of minor morphological features, major periglacial phenomena, such as frost-crack polygons and other types of patterned ground, pingos, rock glaciers, dunes, thermokarst valleys, and collapse features can be easily delineated on vertical aerial photographs at scales of 1:25,000 to 1:30,000. The Large Format Camera (LFC) carried on the October 1984 Space Shuttle mission made more than 2,000 high-resolution (10- to 20-m) black-and-white, color, and false-color infrared photographs of the Earth available to the scientific community (Kreig and others, 1986). Images from the Advanced Spaceborne Thermal Emission and Deflection Radiometer (ASTER) on the Terra satellite and Landsat Enhanced Thematic Mapper Plus (ETM+) images have a 15-m pixel resolution. QuickBird 2 provides images with a pixel resolution of 0.61 m. In the latter example, the increase in spatial resolution of satellite images makes them comparable with the spatial resolution of aerial photographs and hence more useful in mapping periglacial features.

Conventional aerial photography has long been employed to detect and measure changes in permafrost-dominated landscapes. The rates of changes in Arctic coastlines due to the erosion of ice-rich permafrost are well documented by repeated coverage with aerial photographs and satellite images. The Arctic Coastal Dynamics (ACD) Program provides a circum-Arctic assessment of coastline-retreat measurements based on these methods (Rachold and others, 2005; Jorgenson and Brown, 2005) (fig. 37). Kääb (2005) and others provide abundant examples of site-specific investigations of creeping permafrost and



Coastline type	Mean annual erosion rate (ma^{-1})
Bay/Inlet	2.0
Exposed bluffs	2.4
Lagoon	0.7
Tapped basins	1.7
Delta	-2.3

Figure 37.*—Map showing coastline type and mean annual historical erosion rates along the Beaufort Sea coast of Alaska based on analysis of high-resolution imagery and geodetic ground control (Jorgenson and Brown, 2005).*

slope failures in the mountain regions through conventional aerial photography and satellite imagery, and they used differential synthetic aperture radar (SAR) interferometry to map surface deformation of rock glaciers (fig. 38). Earlier studies of rock glaciers in the Andes utilized aerial photography and Landsat imagery (fig. 21) (Corte, 1998). Using Landsat 7 and a segmentation algorithm, Hinkel and others (2005) classified lakes and drained lakes basins of the Arctic Coastal Plain of Alaska based on age, shapes, and orientation (fig. 25).

Even when the pixel resolution of satellite images is lower than that of conventional aerial photographs, satellite images have the advantage of providing regional overviews that require a minimum number of scenes; they also capture repeated and multitemporal or seasonal characteristics of the terrain (including variation in solar-elevation angle above the horizon) (WMO, 2003; Brown and others, 2007). Satellite images increase the opportunities for producing more optimal records of periglacial landforms, especially for recording short-lived features such as icings (figs. 32, 33) and regional progression of soil freezing and thawing (as recorded by freezeup and thaw of lakes).

Duguay and others (2005) reviewed current satellite capabilities for detecting and mapping permafrost and seasonally frozen ground. Among the satellite data readily accessible from international archives are: Landsat Multispectral Scanner (MSS), Thematic Mapper (TM), and Enhanced Thematic Mapper Plus (ETM+) images; images from Satellite pour l'Observation de la Terre (SPOT), and Advanced Spaceborne Thermal Emission Radiometer (ASTER) images. All were, until recently, the most frequently used data sources for interpreting and observing changes in permafrost conditions. IKONOS® (1-m pixels) and QuickBird 2 (0.61-m pixels) satellite images and the declassified CORONA (strategic reconnaissance) satellite photographs (especially from the KH-4 satellite) are providing new capabilities for observing changes and for inferring the presence of permafrost conditions. Landsat satellite imagery of western Siberia was used to detect changes in thaw-lakes hydrology and drainage (Smith and others,

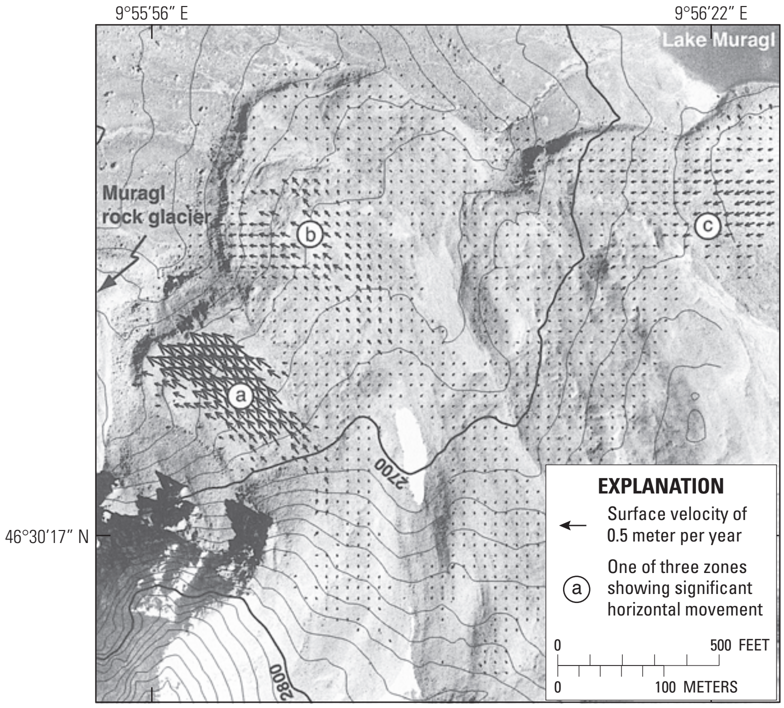


***Figure 38.**—Horizontal surface velocities on part of the Muragl Glacier forefield, Switzerland (from fig. 9-31 in Kääb, 2005).*

2005), and historical photography and IKONOS® images were used to map changes in the extent of thermokarst ponds on the Seward Peninsula of Alaska (Yoshikawa and Hinzman, 2003). Numerous approaches utilize supervised-classification schemes to map categories of active layer thickness, late-summer frozen ground, and vegetation indicators. Images from the pair of Canadian remote-sensing satellites (RADARSAT) have been used to map zones of wintertime heat loss that indirectly approximate the distribution of permafrost. Spaceborne scatterometers and passive-microwave imagery provide data for detecting frozen and unfrozen soils and for mapping seasonal progression of freeze-thaw events over large areas. These and other new satellite-based mapping tools and techniques promise advances in remote-sensing technology for measuring spatial and temporal characteristics of near-surface permafrost and the distribution of seasonally frozen ground (Zhang and others, 2004).

Observational Networks

Long-term, time-series records of permafrost temperatures and seasonal timing of soil thaw at the same locations and across diverse types and regions of terrain are required in order to identify the scale of spatial variation, establish trends, and validate models. Widespread systematic changes in the thickness of the active layer and in the warming of permafrost terrains could have profound effects on the flux of greenhouse gases (especially release of sequestered methane), on the built infrastructure in cold regions, and on landscape processes. It is therefore critical that observational and analytical procedures continue over decadal periods to assess trends and detect cumulative, long-term changes.

Observation of changes in permafrost and in periglacial environments is accomplished by direct field measurements and from remotely sensed data and imagery over varying time intervals. Within the past decade, several polar terrestrial observing networks have been established under the auspices of the International Permafrost Association (IPA) to monitor the rates of change in permafrost and periglacial environments. In 1997, the Global Climate Observing System (GCOS) and the Global Terrestrial Observing System (GTOS) identified the active layer and the thermal state of permafrost as two key cryospheric variables for monitoring in permafrost regions (GCOS, 1997; Burgess and others, 2000). These key parameters are now being monitored under the Global Terrestrial Network for Permafrost (GTN-P) (Smith and Burgess, 2003; Smith and Brown, 2009) through two components—Circumpolar Active Layer Monitoring (CALM) and boreholes (Romanovsky and others, 2002). The Geological Survey of Canada maintains the metadata files for the GTN-P boreholes, including observed data from some of the boreholes [*http://www.gtnp.org*].

The adequacy of the GCOS networks, including GTN-P, is reviewed periodically under GCOS, reporting to the United Nations Framework Convention on Climate Change (WMO, 2003). The 2003 report concluded that "New temperature boreholes and in situ observations of active layer thickness need to be established in both hemispheres." (WMO, 2003, p. 43; see updates in WMO, 2010). The IPA and the Scientific Committee on Antarctic Research and its Expert Group on Permafrost and Periglacial Environments have developed protocols and established a network of boreholes and active layer sites in the Antarctic and subantarctic. Under the auspices of the IPA, collaboration is underway with the World Climate Research Program's Climate and Cryosphere project (WCRP CliC) to develop cooperative programs on the terrestrial cryosphere.

Circumpolar Active Layer Monitoring (CALM)

The CALM network was originally developed in conjunction with the International Tundra Experiment (ITEX) ecological project (Brown and others, 2000). CALM consists of more than 160 sites (fig. 39) in both polar and subpolar regions, and 14 countries participate. Most sites in the network are in Arctic and Subarctic lowlands, although 20 sites are in high-mountain regions of the Northern Hemisphere, at altitudes greater than 1,300 m. A new Antarctic component is being organized and has currently identified 14 sites. CALM represents the only coordinated and standardized program designed to observe and detect decadal changes in the timing of seasonal thawing and freezing of high-latitude soils (Brown and others, 2000; Nelson, 2004a, b; Guglielmin, 2006; Nelson and others, 2008). Although the main purpose

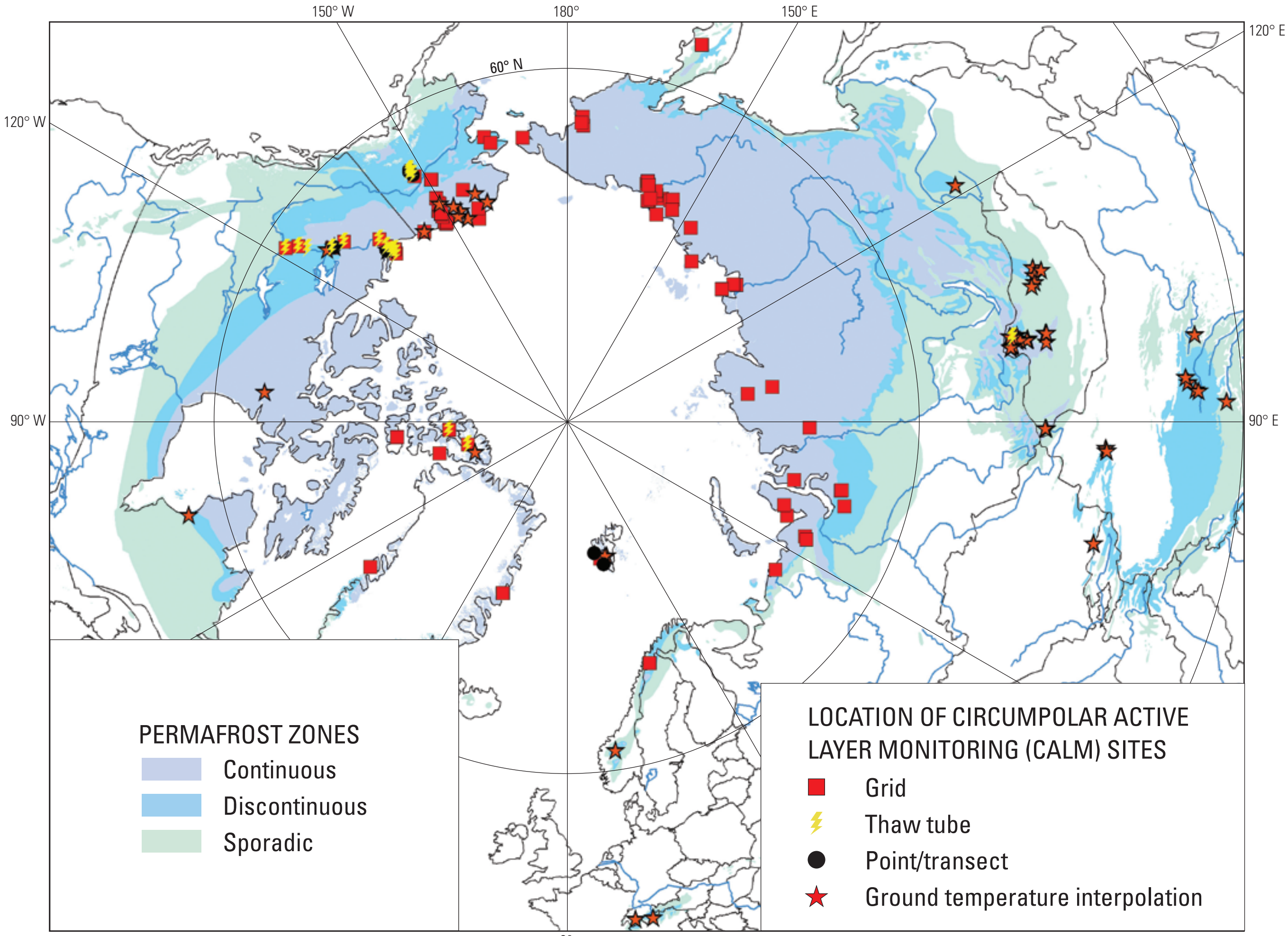


__Figure 39.__—Location of the sites and the type of measurements in the Circumpolar Active Layer Monitoring Network (CALM).

of CALM is to measure the depths of late-summer thaw, many sites include seasonal measurements of soil temperature and moisture and of ground heave and subsidence. Some sites use conventional surveying techniques or differential global positioning systems to ascertain thaw settlement that is due to ground-ice melt and frost heave (Little and others, 2003). Since 2004, the University of Delaware has maintained the CALM database and its associated Web site [*http://www.udel.edu/Geography/calm/*].

Thermal State of Permafrost

The second GTN-P network, designed to record the temperature of permafrost at various depths and time periods, is the Thermal State of Permafrost (TSP) Project (fig. 40) (Brown and Romanovsky, 2008). More than 860 boreholes in 25 countries in continental and mountainous regions of both hemispheres have been identified for new or continued measurements. Many of these boreholes range between 25 and 100 m in depth [*http://nsidc.org/data/g02190.html*].

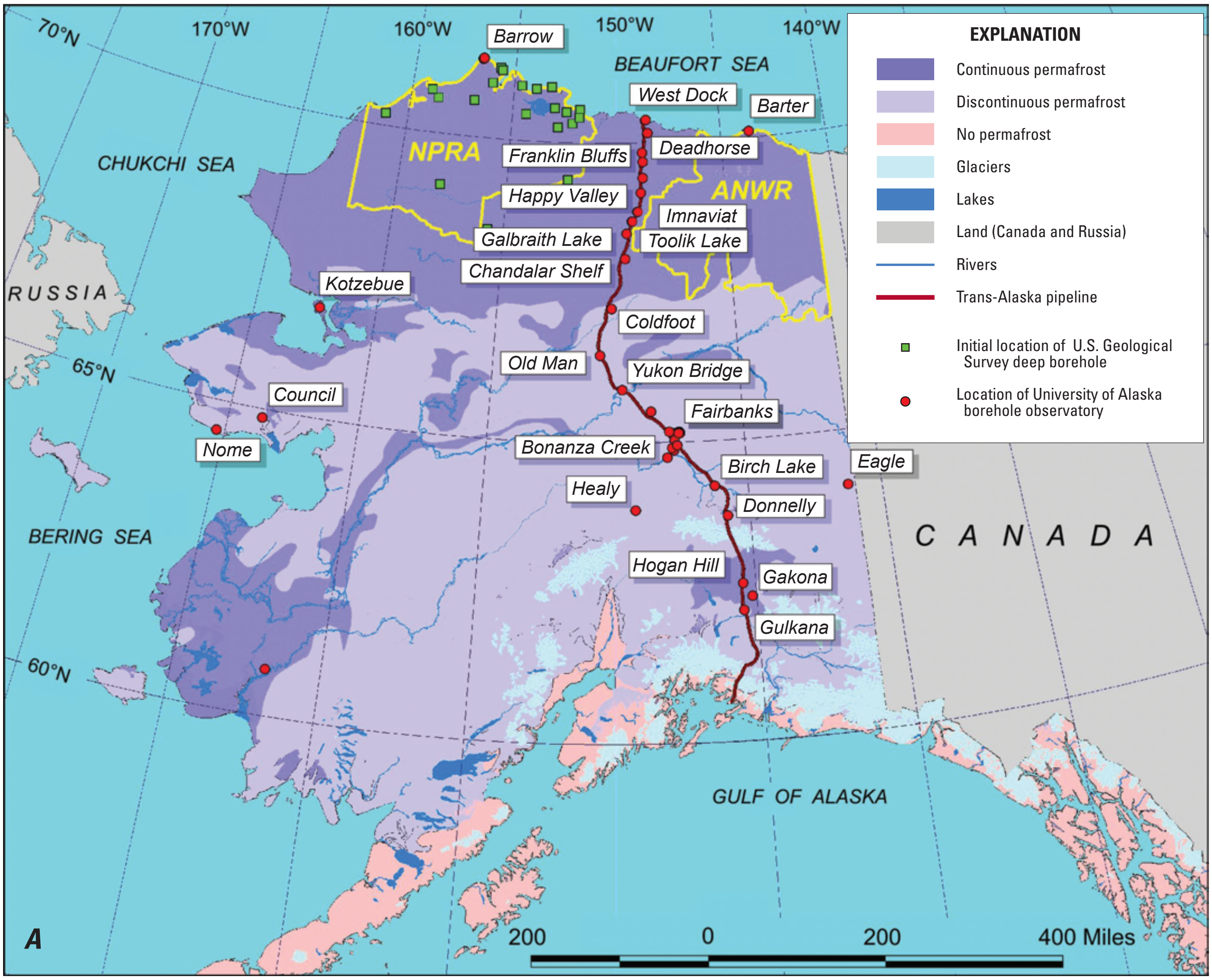
Research groups at the University of Alaska maintain approximately 50 borehole sites, with most of the boreholes ranging from 60 to 80 m in depth (Romanovsky and others, 2002; Osterkamp, 2003a, 2008). The U.S. Geological Survey (USGS) (Lachenbruch, 2002) continues to measure a network of 20 deep boreholes (using abandoned oil wells between 227 to 884 m in depth) in northern Alaska (fig. 41*A*). Figure 41*B* shows warming in °C. at 20-m depth in permafrost, between 1989 and 2007/2008, in the USGS boreholes. Many of these borehole sites have automatic weather stations and soil-temperature measurement probes for estimating site-specific active-layer depths.

In Canada, the Geological Survey of Canada is responsible for coordinating and implementing the framework and infrastructure for a national permafrost monitoring network; it also maintains a network of more than 100 boreholes (typically up to 20 m in depth), established since the mid 1980s, in the Mackenzie Valley and Mackenzie Delta (Smith and others, 2005; Taylor and others, 2006). Université Laval, Québec City, maintains a network of 30 boreholes, up to 20 m in depth, in northern Québec; the network was established in phases between 1979 and 1994.

In Asia, temperatures in thousands of boreholes in Russia and the former Soviet Union were measured during the past 30 to 40 years at research stations and construction sites. These include boreholes in central and southern Yakutia, northern West Siberia, European Russia, and some in East Asia (Romanovsky and others, 2008). Existing boreholes are being monitored in China (Zhao and others, 2004, 2008), Mongolia (Sharkhuu, 2003; Sharkhuu and others, 2007), and Kazakhstan (Marchenko, 1999).

In Europe, the European Union project on Permafrost and Climate in Europe (PACE) initiated a network of cables 100-m long in boreholes from the Alps in the south to Svalbard, Norway, in the North Atlantic (Harris and Haeberli, 2003; Harris and others, 2009). Permafrost Monitoring Switzerland (PERMOS) has a network of 27 boreholes at 13 sites (Vonder Mühll and others, 2008). An expanded Norwegian monitoring program (NORPERM) was developed as part of its International Polar Year program. Figure 9*A–F* illustrates typical time-series from some of these boreholes.

Figure 40.—*Location of borehole sites for the Global Terrestrial Network for Permafrost (GTN-P) and the IPY Thermal State of Permafrost project (provided by Vladimir Romanovsky, University of Alaska Fairbanks).*



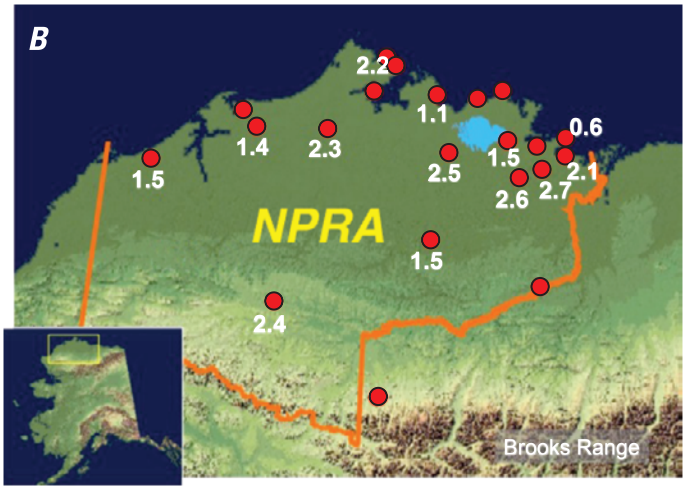


Figure 41.—**A**, *Locations of University of Alaska borehole observatories (provided by Vladimir Romanovsky, University of Alaska Fairbanks) and locations of the U.S. Geological Survey (USGS) deep boreholes in northern Alaska (Lachenbruch and Marshall, 1986); and* **B**, *Warming in degrees Celsius of permafrost at 20-m depth between 1989 and 2007/2008 at USGS borehole sites within and admacent to National Petroleum Reserve in Alaska (NPRA) (courtesy of Gary D. Clow, USGS). On A, NPRA is the National Petroleum Reserve, Alaska; ANWR is Alaska National Wildlife Refuge.*

Arctic Coastal Dynamics

Erosion rates of permafrost-dominated coastlines, particularly those containing massive ground ice, are monitored at key sites on the ground as well as with sequential aerial and high-resolution satellite imagery. The Arctic Coastal Dynamics (ACD) network (Rachold and others, 2005), a joint project of the IPA and the International Arctic Science Committee, established approximately 30 permanent key sites located along the coasts bordering the Arctic Basin (fig. 37). Erosional rates are established by repeated ground measurements or by analysis of sequential aerial photographs. An overall objective of ACD is to estimate the flux of sediments and carbon from coastline erosion to the inner shelf. The ACD is also a regional program of the Land-Ocean Interactions in the Coastal Zone (LOICZ) project of the International Geosphere Biosphere Programme (IGBP). An ACD Web site is maintained at the Alfred Wegener Institute, Potsdam, Germany [*http://www.awi-potsdam.de/acd*].

Periglacial Monitoring

A new periglacial-monitoring program is being established by the IPA Working Group on Periglacial Landforms, Processes, and Climate to measure rates of periglacial processes. Its objective is to achieve a better understanding of geomorphic effects of climatic variations, past, present, and future, by investigating temporal and spatial variability and by promoting standardized field techniques on periglacial processes. The first monitoring site was established in Svalbard, Norway, in 2004 (Matsuoka, 2006).

Network of Permafrost Observatories

As a component of the International Polar Year (IPY), the International Permafrost Association has undertaken the development of an International Network of Permafrost Observatories. When formally implemented, this network of active layer and borehole observatories will include hundreds of sites in both hemispheres. Initially the network involves participants from 25 countries, and it will importantly advance the GTN-P. During the IPY 2007 to 2009, an important first phase resulted in the acquisition of a uniform set of borehole temperatures (snapshot) from all permafrost regions of the Earth. This baseline set of data and a permanent network of observations are part of the IPY-IPA legacy [*http://www.arctic.noaa.gov/reportcard/permafrost.html*].

Permafrost Hazards and Engineering Considerations

Beyond its intrinsic interest as an object of scientific study, the investigation of the distribution and properties of permafrost and ground ice have considerable practical and economic value. Knowledge of permafrost conditions, history, and properties are important for studies of global climatology and of climatic change, for the development of infrastructure for resources, for the exploration and production of oil, gas, and minerals, for slope stability, foundation design, water supply, and waste disposal, and for environmental protection. Since human beings first built in permafrost regions, dwellings and other built structures have been damaged from cumulative subsidence due to the thawing of permafrost but also from heaving due to the expansion of ground ice within the permafrost and from erosion (fig. 42) and solutions have been sought (fig. 43).

Although ground temperatures affect the strength and deformation properties of frozen ground, the presence of excess ground ice is the primary cause of instability of the soil and of loss of ground strength associated with subsequent thawing of ice-rich soil. The loss of ground strength can lead to critical problems for engineered structures, such as highways, railways, pipelines, dams, and building foundations, unless they have been designed and built to withstand these effects. The growth of ground ice, seasonal as well as perennial, and the associated frost heave of the ground can also create problems for design and maintenance, particularly for linear structures such as pipelines, roads, and airfields.

Special precautions in engineering designs are required for construction on permafrost and on ice-rich terrain (Ferrians and others, 1969; Johnston, 1981). Passive techniques include measures to prevent the thawing of the underlying frozen ground, such as elevating buildings and pipelines on piles (figs. 44, 45), placing insulation on the ground surface, and using cooling systems, such as thermosyphons (fig. 46). More aggressive approaches require removing ice-rich soil and replacing it with material that is not susceptible to

Figure 42.—*Failed building as a result of differential thaw of ice-rich permafrost, Fairbanks, Alaska (photograph taken May 2004 by Vladimir Romanovsky, University of Alaska, Fairbanks).*

Figure 43.—*Attempt to control shoreline erosion using riprap at Tuktoyaktuk, Northwest Territories, Canada (photograph by Steve Solomon, Geological Survey of Canada).*

Figure 44.—*Pile foundations to prevent thaw settlement, Russian mining community, Pyramiden, Svalbard, Norway (photograph by Ole Humlum, University of Oslo, Department of Geography and The University Centre in Svalbard).*

Figure 45.—*Trans-Alaska Pipeline System (TAPS) on elevated thermopiles to prevent permafrost thaw (photograph courtesy of Alyeska Pipeline Service Company).*

Figure 46.—*Thermosyphons along the Qinghai-Tibet Railroad, China (photograph courtesy of the Cold and Arid Regions Environmental and Engineering Institute, Chinese Academy of Sciences, Lanzhou, China).*

frost or allowing the excess ice to melt before construction begins (active techniques). Examples of large construction projects using passive techniques are the Trans-Alaska Pipeline (fig. 45) and the new Qinghai-Tibet Railway (fig. 46) (Cheng, 2004, 2005a, b). For all except the smallest structures, however, even foundations designed using passive techniques are likely to be affected by deformation of the underlying ground. Johnston (1981) provided a practical summary of the problems that may develop as a result of the rheological properties of frozen soil.

Cold regions and permafrost engineering have developed into a highly specialized branch of engineering that creatively adapts experience associated with open-pit and underground mines, tailings and mine-waste facilities, highways and roads, airstrips, landfills, dams and water-diversion structures, pipelines, and foundation systems for industrial, military, and commercial uses. Current engineering practice and design also take into account the risks associated with the impact of the globally warming climate on permafrost behavior (US-ARC, 2003; Hayley and Horne, 2006). Natural warming of permafrost in mid-latitude mountainous regions and subsequent melting of the ground ice in bedrock foundations affect ski facilities and slopes. Avalanches such as occurred in the Kolka-Karmodon area, Caucasus Mountains, North Ossetiya, Russia (fig. 47) (Kotlyakov and others, 2010), and rock slides triggered by permafrost thaw in rock walls, such as occurred on the Matterhorn in July 2003, and in the Thurnweiser (Italy) in 2004, are becoming increasingly common (Kääb, 2005).

In addition to the International Conferences on Permafrost (Brown and Walker, 2007), other engineering and geotechnical conferences are held frequently with participants reviewing the current state of knowledge and best engineering practices. These conferences and publications include the Cold Regions Specialty conferences of the American Society of Civil Engineers (Zufelt, 1999), the Canadian Geotechnical Society, and various cold-regions and geotechnical journals, such as Journal of Cold Regions Engineering, Cold Regions Science and Technology, Permafrost and Periglacial Processes, and so on.

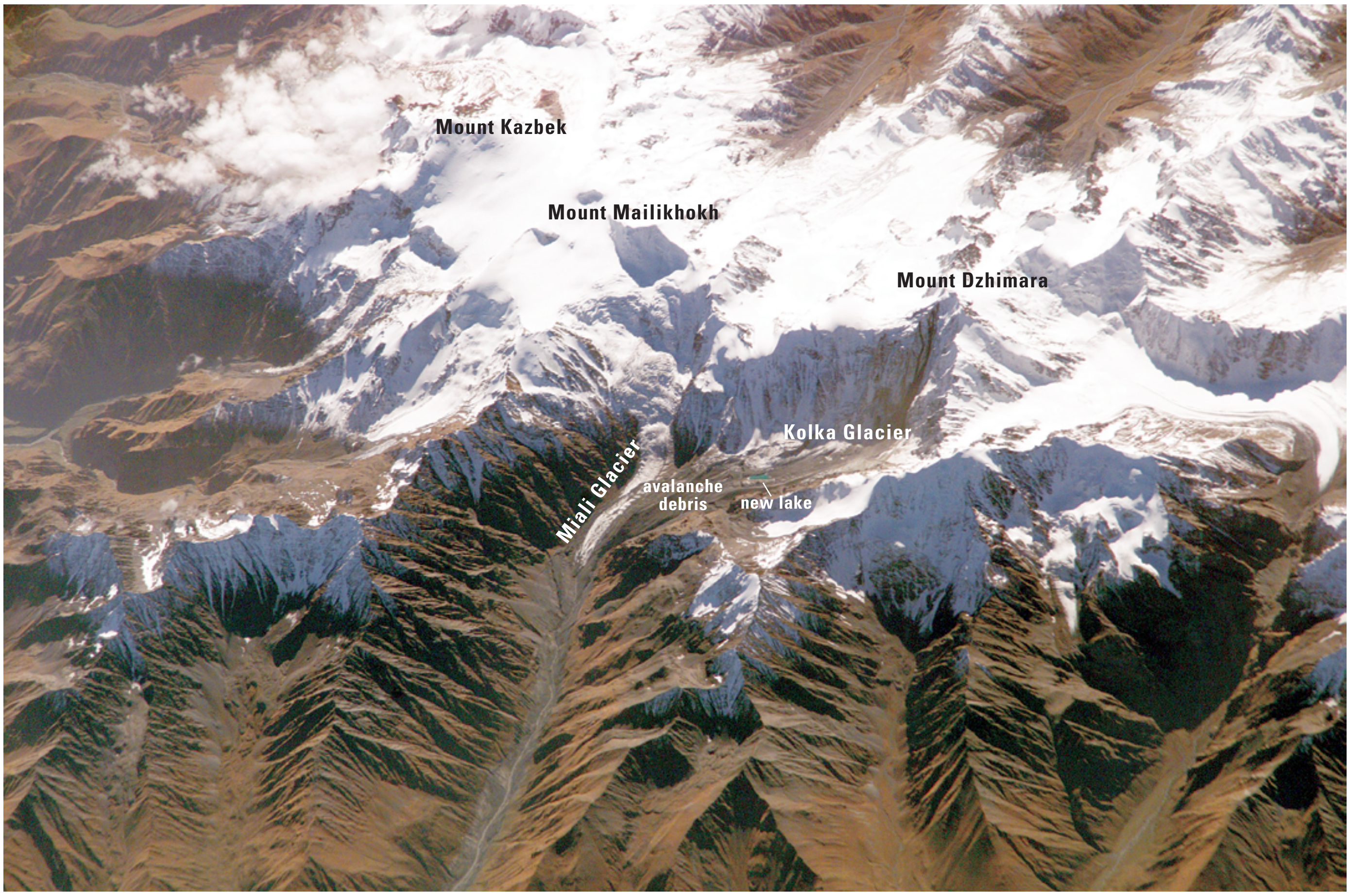


***Figure 47.**—Kolka-Karmadon ice-rock avalanche of 20 September 2002. NASA International Space Station photograph no. ISS005-E-17830 (photograph courtesy of Office of Public Affairs, National Aeronautics and Space Administration, Washington, D.C.)*

Change Scenarios for the 21st Century

Changes in climate have short-term and long-term effects on near-surface permafrost and are mediated by changes in the surface-boundary layer (snow, vegetation, and soil covers) (Zhang and others, 2007). Because permafrost is three dimensional, changes can occur in areal and vertical (depth or thickness) extent (figs. 2, 3, 6). During the Little Ice Age, permafrost cooled and expanded in areal and volumetric extent in mountains and in regions of discontinuous permafrost. However, the timing of the Little Ice Age varies somewhat from region to region on the planet, but the period 1300 to 1900 C.E. is generally accepted.

Deepening or thickening of the active layer or increases in the number and extent of taliks do not everywhere result in the short-term disappearance of permafrost. Even permafrost close to the thawing point can persist for many centuries or millennia (fig. 3, relict zone in West Siberia). The actual response of the terrain at a specific locality is difficult to forecast, because of the complex interactions of climate, vegetation, snow cover, and soil conditions, as well as the existing ground-ice regime.

Over long periods of time, climate warming will lead to a thickening of the active layer, loss of the uppermost ice-rich layer of permafrost, increased talik formation, and a reduction in the areal extent of near-surface permafrost in the warmer discontinuous permafrost zone. Frozen peatlands and palsa will undergo degradation—easily visible by surface changes. In areas with high ground-ice content and extensive thermokarst, widespread slope failures can be expected to develop. This loss of ground ice and of any buried glacial ice will be irreversible, in that, even with a much colder climate, it would take centuries to millennia for comparable quantities of ground ice to redevelop. Small-scale maps that depict changes in the boundaries of the permafrost zones during a 50- to 100-year period are generally misleading, because they do not consider the great variability in frozen and unfrozen terrain and the resistance to thaw at depth.

Thawing of ice-rich permafrost, regardless of its location, has a considerable effect on the landscape and natural ecosystems and on human social and economic activities. Erosion or subsidence of ice-rich landscapes, alterations to hydrological processes, and release of greenhouse gases to the atmosphere will occur. The stability of buildings, linear structures, including roads and pipelines, and communication links will be affected. The observed and forecast changes in permafrost stress the necessity to monitor permafrost dynamics on the ground (particularly temperatures) and by remote sensing for timely assessment and forecasts of the possible negative effects of permafrost degradation on ecosystems and built infrastructures.

In "Climate Change," the Second Assessment of the Intergovernmental Panel on Climate Change, the authors concluded that most of the ice-rich discontinuous permafrost could disappear over the span of a century (Houghton and others, 1996; Watson and others, 1996). The Third IPCC report "Climate Change 2001: Impacts, Adaptation, and Vulnerability" (McCarthy and others, 2001) recognized that thickening of the active layer is expected and that modeling studies indicate that large areas of permafrost terrain will begin to thaw (Anisimov and others, 2001). The Arctic Climate Impact Assessment (ACIA, 2004, 2005) concluded that permafrost degradation is likely to occur over 10 to 20 percent of the present permafrost area, and the southern limit of permafrost is likely to move northward by several hundred kilometers (Walsh and others, 2005). The Fourth IPCC report "Climate Change 2007: The Physical Science Basis" (Lemke and others, 2007) refined estimates of active-layer changes and the extent of additional permafrost degradation. Although recent increases in active-layer thickness are observed in response to seasonal fluctuations in climate, overall hemispherical changes in the active layer as observed from the CALM network (fig. 39) are not conclusive. Some recent studies suggest that the increase in permafrost temperatures may be slowing (Richter-Menge and others, 2007).

Recent studies have revealed that active permafrost degradation is occurring in Alaska (Jorgenson and others, 2001), Canada (Camill, 2005; Zhang and others, 2006), Mongolia (Sharkhuu, 2003), China (Zhao and others, 2004), and Svalbard and Scandinavia (Isaksen and others, 2007). Pavlov and Malkova (2005) conclude that for Russia, during the last 30 to 35 years, a general tendency toward warming is observed in permafrost temperatures, although with distinct regional differences.

At present, no coupled climate model exists that would take all changes into account. However, by choosing a future climate scenario and assuming certain changes in vegetation and hydrology or both, it is possible to apply these driving forces to a permafrost model in order to forecast future permafrost dynamics on a regional or even on a circumpolar scale. Figure 48 shows an example of such a projection for the entire Northern Hemisphere permafrost domain based on a spatially distributed permafrost model developed at the Permafrost Laboratory, Geophysical Institute, University of Alaska Fairbanks (Sazonova and Romanovsky, 2003; Sazonova and others, 2004; Romanovsky, Gruber, and others, 2007; Romanovsky, Sazonova, and others, 2007). Three time intervals are considered: the first represents conditions in 2000 (fig. 48*A*), changes in permafrost that may occur by 2050 (fig. 48*B*), and the end of the 21st century (fig. 48*C*). According to this specific climate scenario, the modeling results show that permafrost will be actively thawing within the area in 2000 occupied by discontinuous permafrost with temperatures between 0° and -2°C.

Northern soils, peatlands, and frozen organic rich sediments contain large amounts of organic carbon (fig. 49). In a recent report by Tarnocai and others (2009), carbon pools were estimated to be 191.29 Pg for the 0 to 30-cm depth, 495.80 Pg for the 0 to 100-cm depth, and 1,024.00 Pg for the 0–300-cm depth. Carbon pools in layers deeper than 300 cm were estimated to be 407 Pg for edoma deposits and 241 Pg in deltaic deposits. In total, the northern permafrost region contains approximately 1,672 Pg of organic carbon, of which approximately 1,466 Pg or 88 percent, occurs in perennially frozen soils and deposits. This 1,672 Pg of organic carbon would account for approximately 50 percent of the estimated global belowground organic carbon pool. Deep, perennially frozen sediments, both onshore and beneath the Arctic shelves, contain methane hydrates. Recent reports demonstrate that active thermokarst and thermal erosion releases organic matter from thawing permafrost into anaerobic lake bottoms, resulting in exceptionally high rates of production and emission of methane (Walter and others, 2006; Zimov and others, 2006) (fig. 50). Thawing of organic-rich permafrost and the expansion of existing thermokarst lakes would enhance regional emissions, resulting in positive feedbacks to climate warming (Smith and others, 2005; Walter and others, 2006, 2007; Anthony, 2009). If voluminous quantities of permafrost thaw, many tens of billions of metric tons of methane could be emitted from eroding lakes (Walter and others, 2006).

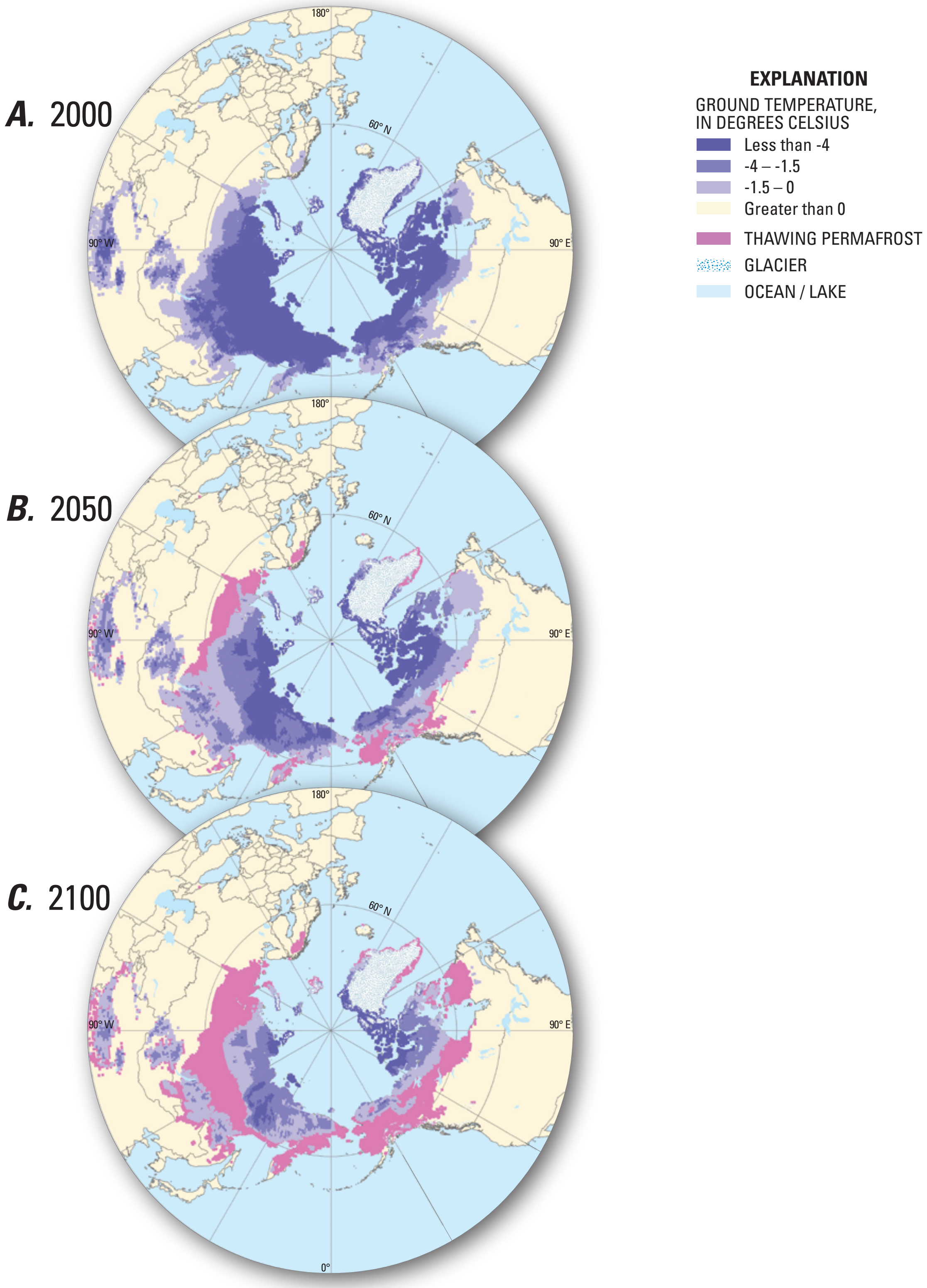


Figure 48.*—Modeled circumpolar permafrost temperatures (mean annual temperature at the permafrost surface) for* **A***, 2000;* **B***, 2050, and* **C***, 2100 (modified from Romanovsky, Gruber, and others, 2007).*

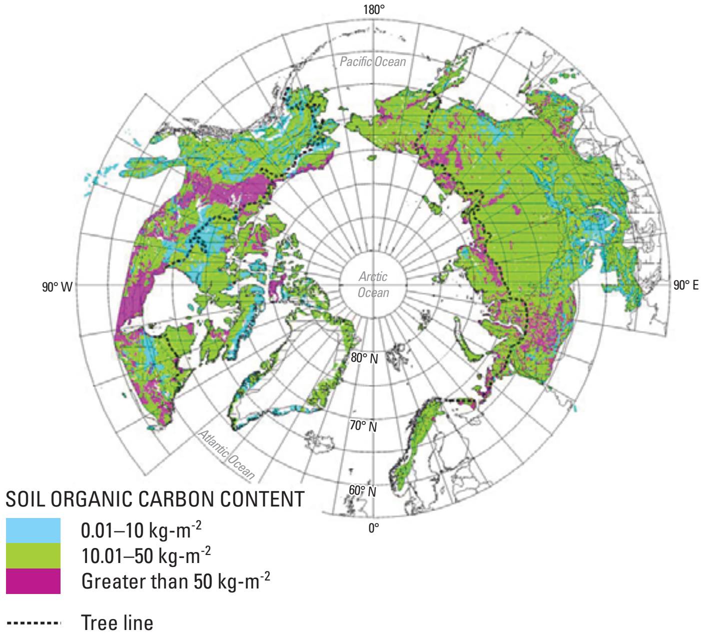


Figure 49.—*Distribution of soil organic carbon content in the northern circumpolar permafrost region based on data in the Northern Circumpolar Soil Carbon Database (NCSCD) (Tarnocai and others, 2009).*

Figure 50.—*Combustion of methane from organic-rich sediments at Shuchi Lake, Siberia, Russia, in March 2007. Katey W. Anthony is on the left, Nikita Zimov on the right. Photograph by Sergey A. Zimov, Director, Northeast Science Station, Cherskii, Republic of Sakha (Yakutia), Russia. (Photograph courtesy of Katey W. Anthony, University of Alaska Fairbanks, Alaska.)*

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