11 THE FUTURE: THE CRYOSPHERE

The nature of the cryosphere

The cryosphere is that part of Earth's surface or subsurface environment that is composed of water in the solid state. It includes snow, sea ice, the polar ice sheets, mountain glaciers, river and lake ice, and permafrost (permanently frozen subsoil).

The cryosphere contains nearly 80% of all Earth's freshwater. Perennial ice covers about 11% of Earth's land surface and 7% of the world's oceans, while permafrost underlies about 25% of it. Seasonal snow has the largest area of any component of the global land surface; at its maximum in late winter it covers almost 50% of the land surface of the Northern Hemisphere.

Good background studies on the cryosphere include Knight (1999), and Benn and Evans (1998) on glaciation, and French (1996), Clark (1988) and Washburn (1979) on permafrost and periglacial environments.

Ice sheets are ice masses that cover more than 50,000 km^2. The Antarctic ice sheet covers a continent that is a third bigger than Europe or Canada and twice as big as Australia. It attains a thickness that can be greater than 4000 m, thereby inundating entire mountain ranges. The Greenland ice sheet only contains 8% of the world's freshwater ice (Antarctica has 91%), but nevertheless covers an area 10 times that of the British Isles. The Greenland ice fills a huge basin that is rimmed by ranges of mountains, and has depressed Earth's crust beneath.

The Antarctic ice sheet (Figure 11.1) is bounded over almost half of its extent by ice shelves. These are floating ice sheets nourished by the seaward extensions of the land-based glaciers or ice streams and by the accumulation of snow on their upper surfaces. Ice-shelf thicknesses vary, and the seaward edge may be in the form of an ice cliff up to 50 m above sea level with 100–600 m below. At its landward edge the Ross Ice Shelf is 1000 m thick. It covers an area greater than that of California.

Ice caps have areas that are less than 50,000 km^2 but still bury the landscape. The world's alpine or mountain glaciers are numerous, and in all there are probably over 160,000 on the face of the earth. Their total surface area is around 530,000 km^2.

Permafrost underlies large expanses of Siberia, Canada, Alaska, Greenland, Spitzbergen, and north-

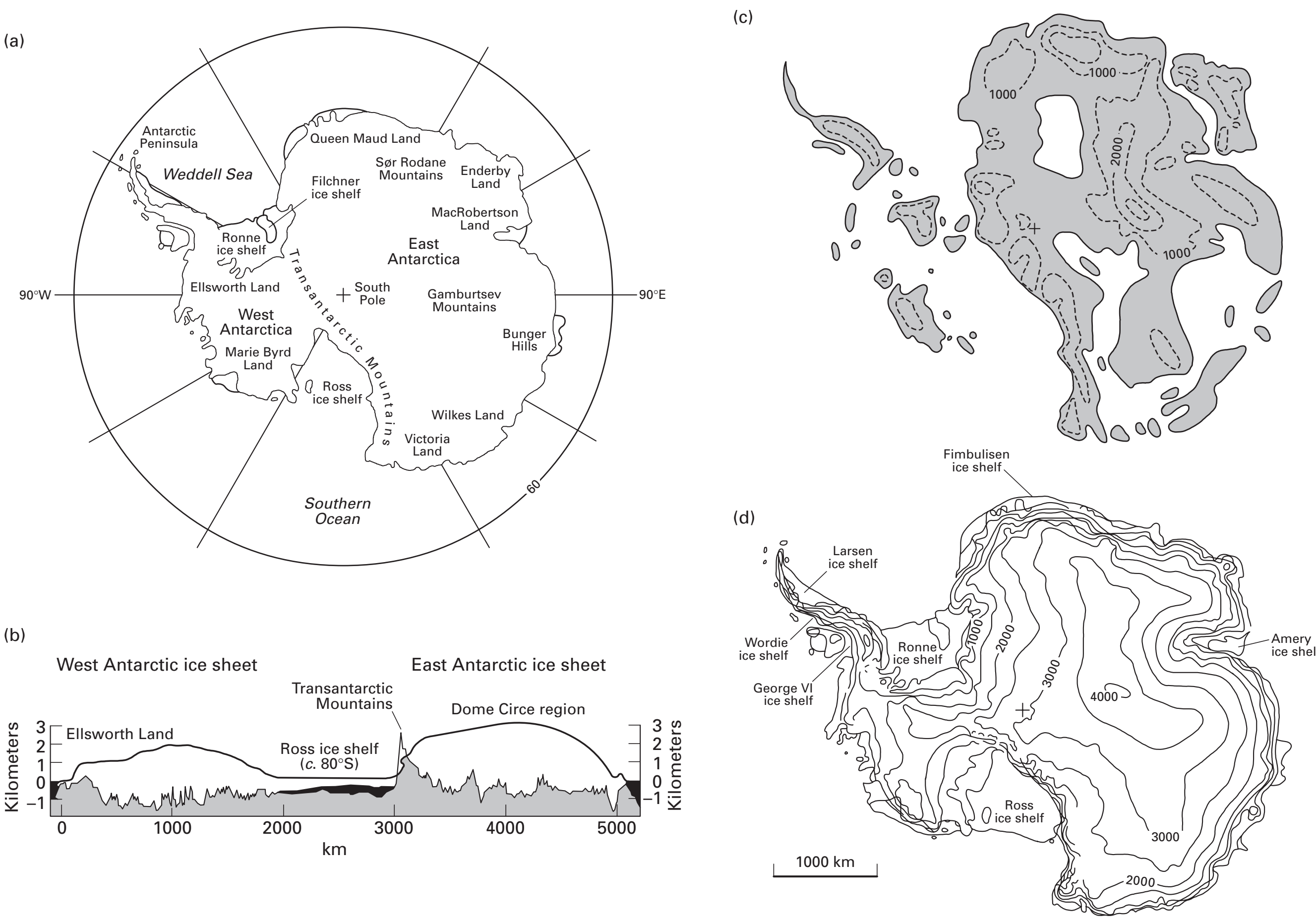

Figure 11.1 The Antarctic ice sheets and shelves. (a) Location map of Antarctica. (b) Cross-section through the East and West Antarctic ice sheets, showing the irregular nature of the bedrock surface, ice thickness, and the floating ice shelves. (c) Subglacial relief (in meters) and sea level. The white areas are below sea level. (d) Surface elevations on the ice sheet in meters.

western Scandinavia (Figure 11.2). Permafrost also exists offshore, particularly in the Beaufort Sea of the western Arctic and in the Laptev and East Siberian Seas, and at high elevation in mid-latitudes, such as the Rocky Mountains of North America and the interior plateaux of central Asia. It occurs not only in the tundra and polar desert environments poleward of the tree line, but also in extensive areas of the boreal forest and forest-tundra environments.

Above the layer of permanently frozen ground there is usually a layer of soil in which temperature conditions vary seasonally, so that thawing occurs when temperatures rise sufficiently in summer but freeze in winter or on cold nights. This zone of freeze–thaw processes is called the *active layer*. It varies in thickness, ranging from 5 m where unprotected by vegetation to typical values of 15 cm in peat.

Conventionally, two main belts of permafrost are identified. The first is the zone of continuous permafrost; in this area permafrost is present at all localities except for localized thawed zones, or taliks, existing beneath lakes, river channels and other large water bodies which do not freeze to their bottoms in winter. In the discontinuous permafrost zone small-scattered unfrozen areas appear.

Maximum known depths of permafrost reach 1400–1450 m in northern Russia and 700 m in the north of Canada, regions of intense winter cold, short cool

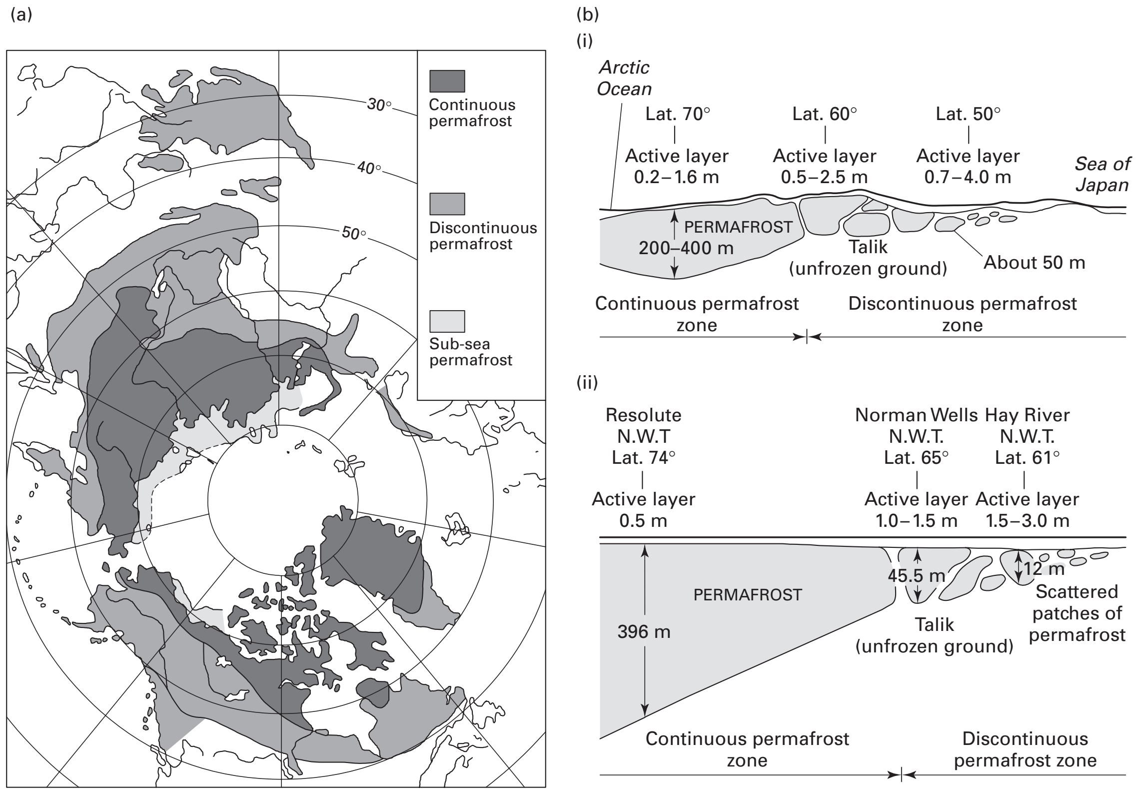

Figure 11.2 (a) The distribution of the main permafrost types in the Northern Hemisphere. (b) Vertical distribution of permafrost and active zones in longitudinal transects through (i) Eurasia and (ii) northern America.

summers, minimal vegetation, and limited snowfall. In general, the thickness decreases equatorwards. Sporadic permafrost tends to occur between the –1°C and –4°C mean annual air temperature isotherms, while continuous permafrost tends to occur to the north of the –7°C to –8°C isotherm.

The various components of the cryosphere are geomorphologically highly important. Glaciers and ice sheets not only produce their own suites of erosional and depositional landforms, but also have numerous indirect effects on phenomena such as sea levels, river flows, and loess (eolian silt deposits). Likewise, snow has direct geomorphologic effects, termed nivation, but also has great significance for streamflow regimes. Frost is an important cause of rock weathering and subsurface permafrost is fundamental in terms of such phenomena as patterned ground, slope stability, runoff, land subsidence, and river and coast erosion. Sea ice also plays a major role through its influence on wave activity level along cold coastlines.

Because of the obvious role of temperature change in controlling the change of state of water to and from the liquid and solid sates, global warming has the potential to cause very major changes in the state of the cryosphere.

The polar ice sheets

Three main consequences of warming may be discerned for ice sheets (Drewry, 1991): ice temperature rise and attendant ice flow changes; enhanced basal melting beneath ice shelves and related dynamic response; and changes in mass balance.

Temperatures of the ice sheets will rise due to the transfer of heat from the atmosphere above. However, because of the slow vertical conduction of heat through the thick ice column, the timescale for this process is relatively long (10^2–10^3 years). As the ice warms, it softens and can undergo enhanced deformation, which could increase the discharge of ice into the oceans. However, Drewry argues that it can be discounted as a major factor on a timescale of decades to a century.

Ice shelves would have enhanced basal melt rates if sea-surface temperatures were to rise. This could lead to thinning and weakening of ice shelves (Warner and Budd, 1990). Combined with reduced underpinning from grounding points as sea level rose, this would result in a reduction of backpressure on ice flowing from inland. Ice discharge through ice streams might therefore increase. There are some studies (e.g., De Angelis and Skvarca, 2003) that have found evidence of ice streams increasing their velocities when ice shelves have collapsed, but this is not invariably the case (Vaughan and Doake, 1996). In a wide ranging review, Bennett (2003) has assembled evidence that although ice streams have high velocities and are very dynamic components of the ice sheet environment, and also display highly variable mass balances, collectively they appear to be locally in balance. Indeed, some workers have identified a positive mass balance (Joughlin and Tulaczyk, 2002).

The West Antarctic Ice Sheet (WAIS), because it is a marine ice sheet that rests on a bed well below sea level, is likely to be much more unstable than the East Antarctic Ice Sheet, from which it is separated by the Transantarctic Mountains. If the entire WAIS were discharged into the ocean, the sea level would rise by 5 or 6 m. Fears have been expressed over many years that it could be subject to rapid collapse, with disastrous consequences for coastal regions all over the world.

One extreme view was postulated by Mercer (1978), who believed that predicted increases in temperature at 80°S would start a 'catastrophic' deglaciation of the area, leading to a sudden 5 m rise in sea level. A less extreme view was put forward by Thomas et al. (1979). They recognized that higher temperatures will weaken the ice sheets by thinning them, enhancing lines of weakness, and promoting calving, but they contended that deglaciation would be rapid rather than catastrophic, the whole process taking 400 years or so. Robin (1986) took a broadly intermediate position. He contended (p. 355) that:

> A catastrophic collapse of the West Antarctic ice sheet is not imminent, but better oceanographic knowledge is required before we can assess whether a global temperature rise of 3.5°C might start such a collapse by the end of the next century. Even then it is likely to take at least 200 years to raise sea level by another 5 m.

The WAIS has part of its ice grounded on land below sea level, and part in the form of floating extensions called ice shelves that move seaward but are confined horizontally by the rocky coast. The boundary between grounded and floating ice is called the grounding line (Oppenheimer, 1998). In 1974, Weertman suggested that very rapid grounding line retreat could take place in the absence of ice-shelf backpressure. He demonstrated (see Figure 11.3) that if the bedrock slopes downward in the inland direction away from the grounding line, as is the case with some of the WAIS ice streams, then if an ice shelf providing buttressing becomes unpinned owing to either melting or global sea-level rise, the grounded ice would accelerate its flow, become thinner and rapidly float off its bed.

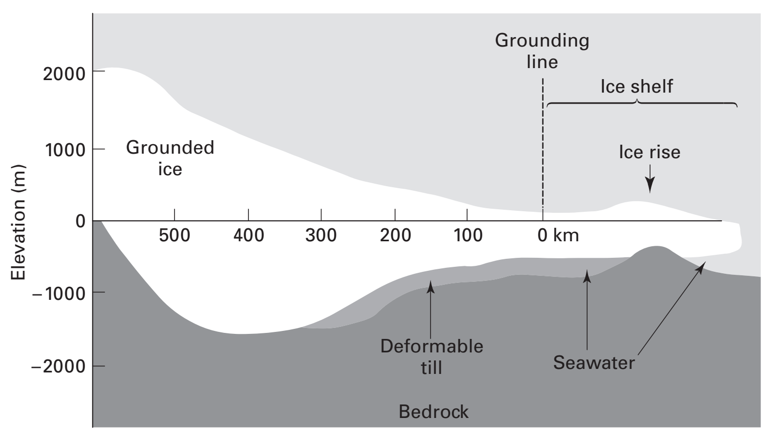

Figure 11.3 Cross-section of an ice stream and ice shelf of a marine ice sheet, indicating location of grounding line, bedrock rise on the ocean floor, and possible extent of deformable till. The thickness of the till layer, actually a few meters, is exaggerated for clarity. Sea-level rise due to collapse of the West Antarctic Ice Sheet (WAIS) was estimated by T. Hughes. (Source: Oppenheimer, 1998, figure 2, p. 327 in *Nature*, volume 393, 28 May 1998 © Macmillan Publishers Ltd 1998.)

Whether the WAIS will collapse catastrophically is still being debated (Oppenheimer, 1998), but there is some consensus that the WAIS will most probably not collapse in the next few centuries (Vaughan and Spouge, 2002). Bentley (1997) is amongst those who have expressed doubts about the imminence of collapse, believing 'that a rapid rise in sea level in the next century or two from a West Antarctic cause could only occur if a natural (not induced) collapse of the WAIS is imminent, the chances of which, based on the concept of a randomly timed collapse on the average of once every 100,000 years, are on the order of 0.1%' (see also Bentley, 1998).

Greenland and Antarctica will react very differently to global warming. This is in part because Antarctica, particularly in its interior, is very cold indeed. This means that the moisture content of the atmosphere over it is very low, ultimately suppressing the amount of snow and ice accumulation that can occur. An increase in temperature may therefore lead to more precipitation and greater accumulation of ice. With Greenland, the situation is different. Whereas in Figure 11.4 Antarctica lies to the left of the mass balance maximum, much of Greenland lies to the right of the maximum so that the total surface mass balance will decrease in the event of global warming (Oerlemans, 1993: 153). Indeed, Gregory et al. (2004) have argued that if the average temperature in Greenland increases by more than about 3°C the ice sheet would be eliminated, causing the global average sea-level to rise by 7 m over a period of 1000 years or more.

However, the state of the Greenland ice sheet does not solely depend on the rate of surface ablation. Flow of the ice is enhanced by rapid migration of surface meltwater to the ice–bedrock interface, and studies have shown that ice accelerates in summer (when melt occurs), that there is a near coincidence of ice acceleration with the duration of surface melting, and that interannual variations in ice acceleration are correlated with variations in the intensity of surface melting. This coupling between surface melting and ice-sheet flow provides a mechanism for rapid response of ice sheets to warming (Zwally et al., 2002; Parizek and Alley, 2004). Some confirmation of this is provided by laser altimeter surveys, which show rapid ice thinning below 1500 m (Krabill et al., 1999; Rignot and Thomas, 2002). They argue that the glaciers are thinning as a result of increased creep rates brought about by decreased basal friction consequent upon water penetrating to the bed of the glacier. They see this as a mechanism for transfer of ice-sheet mass to the oceans that is potentially larger than could be achieved by surface melting alone.

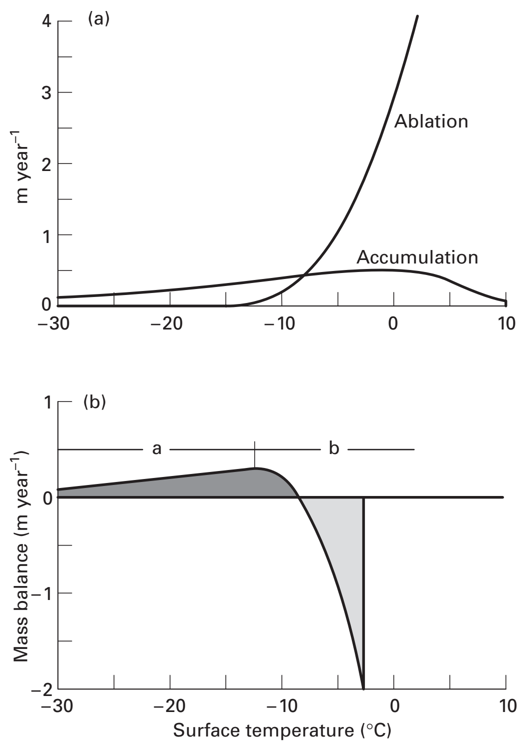

Figure 11.4 (a) Generalized curve of ablation and accumulation in relation to surface temperature. In warmer areas, such as Greenland, an increase in temperature may lead to rapid increases in ablation. (b) Generalized curve of the mass balance of ice sheets in relation to surface temperature. In the very cold environment of Antarctica, an increase in temperature may lead to increased accumulation of snow and ice and hence to a positive mass balance, whereas in the warmer environment of Greenland, increased rates of ablation may lead to a negative mass balance. (Source: Oerlemans, 1993, figure 9.4.)

Valley glaciers and small ice caps

Together the present Greenland and Antarctic ice sheets contain enough water to raise sea level by almost 70 m. By contrast, valley glaciers and small ice caps contain only a small amount – *c.* 0.5 m of sea-level equivalent. However, because of their relatively short response times, associated with large mass turnover, they contribute significantly to sea-level fluctuations on a century timescale. Indeed, there is a clear observational record that valley glaciers are sensitive to climate change, and we have many examples of changes in glacier length of several kilometers since the end of the Little Ice Age (Oerlemans et al., 1998).

Since the nineteenth century, many of the world's alpine glaciers have retreated up their valleys as a consequence of the climatic changes, especially warming, that have occurred in the past 100 or so years since the ending of the Little Ice Age (Oerlemans, 1994). Studies of the changes of snout positions obtained from cartographic, photogrammetric, and other data therefore permit estimates to be made of the rate at which retreat can occur (Figure 11.5). The rate has not been constant, or the process uninterrupted. Indeed, some glaciers have shown a tendency to advance for some of the period. However, if one takes those glaciers that have shown a tendency for a fairly general retreat (Table 11.1) it becomes evident that as with most geomorphologic phenomena there is a wide range of values, the variability of which is probably related to such variables as topography, slope, size, altitude,

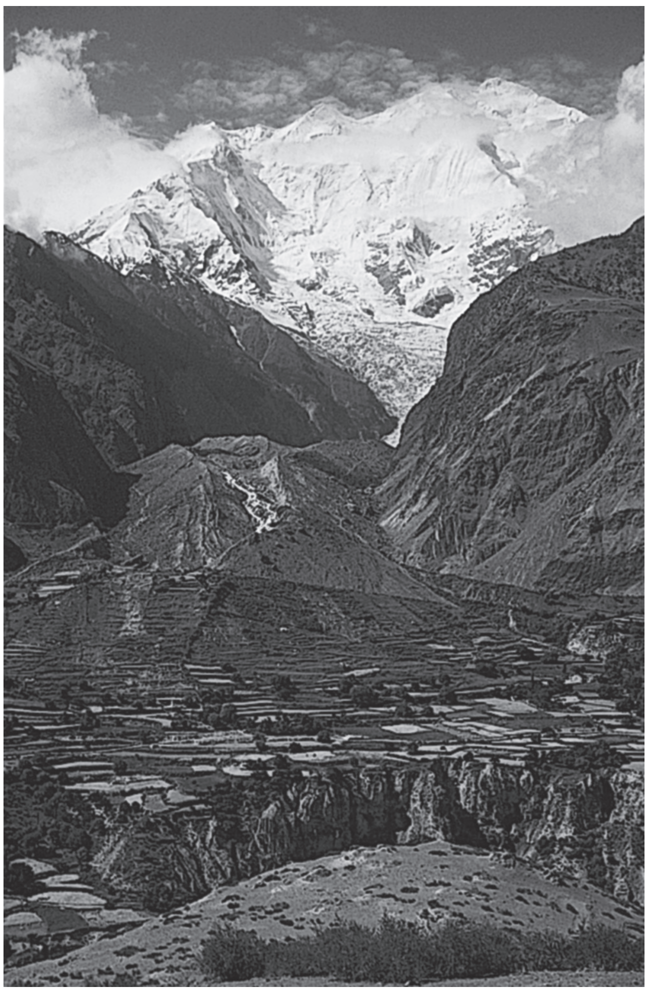

Figure 11.5 The Minapin Glacier snout in the Karakoram Mountains of Hunza, Pakistan. Behind the cultivated fields in the middle distance is a mass of moraine, which was deposited by the glacier in the Little Ice Age. The glacier has now retreated up its valley.

Table 11.1 Retreat of glaciers in meters per year in the twentieth century. Source: tables, maps, and text in Grove (1988)

(a) Individual glaciers

Location	*Period*	*Rate*
Breidamerkurjökull, Iceland	1903–48	30–40
	1945–65	53–62
	1965–80	48–70
Lemon Creek, Alaska	1902–19	4.4
	1919–29	7.5
	1929–48	32.9
	1948–58	37.5
Humo Glacier, Argentina	1914–82	60.4
Franz Josef, New Zealand	1909–65	40.2
Nigardsbreen, Norway	1900–70	26.1
Austersdalbreen, Norway	–	21
Abrekkbreen, Norway	–	17.7
Brikdalbreen, Norway	–	11.4
Tunsbergdalsbreen, Norway	–	11.4
Argentière, Mont Blanc, France	1900–70	12.1
Bossons, Mont Blanc France	1900–70	6.4
Oztal Group, Switzerland	1910–80	3.6–12.9
Grosser Aletsch, Switzerland	1900–80	52.5
Carstenz, New Guinea	1936–74	26.2

(b) Region/country. Source: Oerlemans (1994)

Region	*Period*	*Mean rate*
Rocky Mountains	1890–1974	15.2
Spitzbergen	1906–1990	51.7
Iceland	1850–1965	12.2
Norway	1850–1990	28.7
Alps	1850–1988	15.6
Central Asia	1874–1980	9.9
Irian Jaya	1936–1990	25.9
Kenya	1893–1987	4.8
New Zealand	1894–1990	25.9

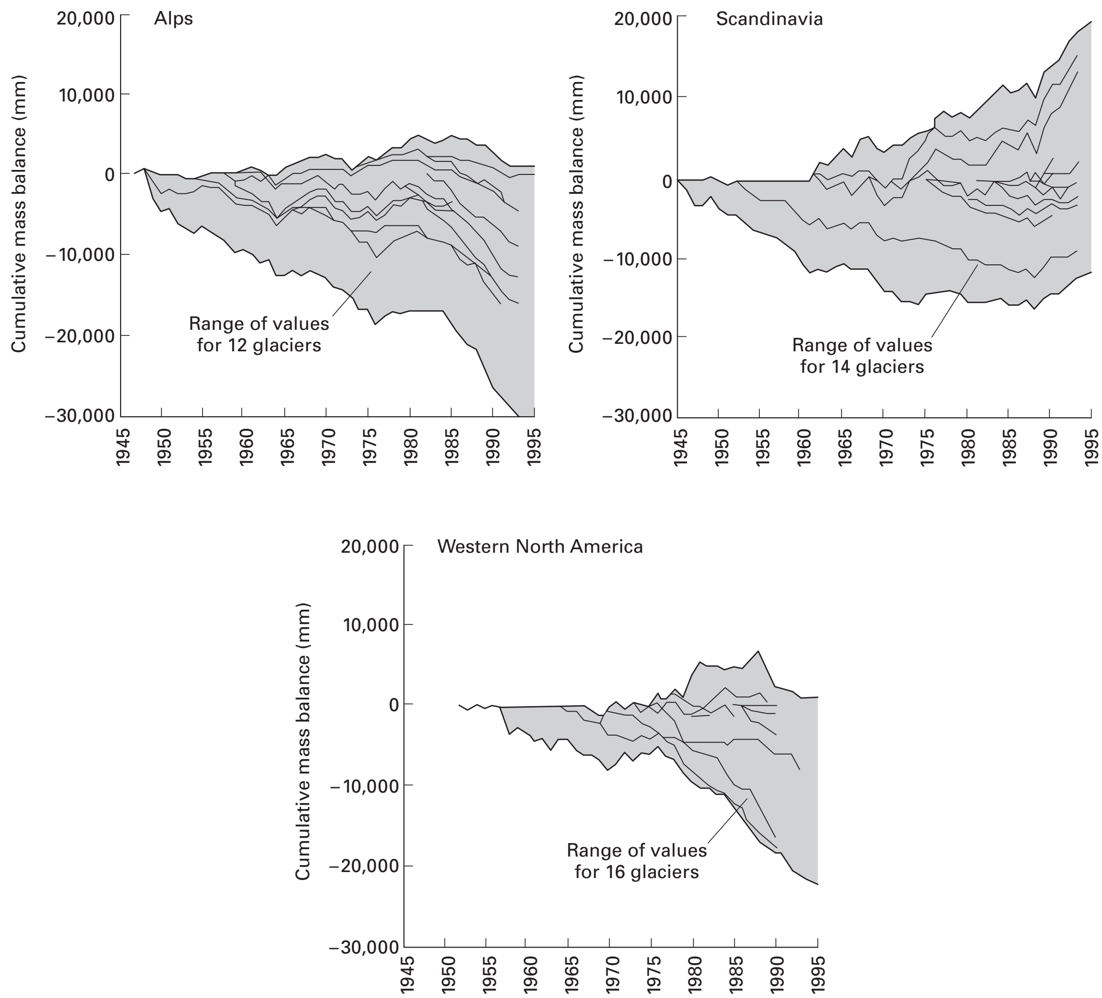

Figure 11.6 Cumulative mass balance for the Alps (top left), Scandinavia (top right), and western North America (bottom).

accumulation rate, and ablation rate. It is also evident, however, that rates of retreat can often be very high, being of the order of 20 to 70 m per year over extended periods of some decades in the case of the more active examples. It is therefore not unusual to find that over the past 100 or so years alpine glaciers in many areas have managed to retreat by some kilometers.

Fitzharris (1996: 246) suggested that since the end of the Little Ice Age, the glaciers of the European Alps have lost about 30 to 40% of their surface area and about 50% of their ice volume. In Alaska (Arendt et al., 2002), glaciers appear to be thinning at an accelerating rate, which in the late 1990s amounted to 1.8 m per year. In China, monsoonal temperate glaciers have lost an amount equivalent to 30% of their modern glacier area since the maximum of the Little Ice Age. Not all glaciers have retreated in recent decades (Figure 11.6) and current glacier tendencies for selected regions are shown. These are expressed in terms of their mass balance, which is defined as the difference between gains and losses (expressed in terms of water equivalent). In the European Alps (top left) a general trend toward mass loss, with some interruptions in the mid-1960s, late 1970s, and early 1980s, is observed. In Scandinavia (top right), glaciers close to the sea have seen a very strong mass gain since the 1970s, but mass losses have occurred with the more continental glaciers. The mass gain in western Scandinavia could be explained by an increase in precipitation, which more than compensates for an increase in ablation caused by rising temperatures. Western North America (bottom) shows a general mass loss near the coast and in the Cascade Mountains (Hoelzle and Trindler, 1998).

The positive mass balance (and advance) of some Scandinavian glaciers in recent decades, notwithstanding rising temperatures, has been attributed to increased storm activity and precipitation inputs coincident with a high index of the NAO in winter months since 1980 (Nesje et al., 2000; Zeeberg and Forman, 2001). In the case of Nigardsbreen (Norway), there is a strong correlation between mass balance and the NAO index (Reichert et al., 2001). A positive mass balance phase in the Austrian Alps between 1965 and 1981 has been correlated with a negative NAO index (Schoner et al., 2000). Indeed, the mass balances of glaciers in the north and south of Europe are inversely correlated (Six et al., 2001).

Glaciers that calve into water can show especially fast rates of retreat. Indeed, retreat rates of greater than 1 km per year are possible (Venteris, 1999). In Patagonia, in the 1990s, retreat rates of up to 500 m per year were observed. This rapid retreat is accomplished by iceberg calving, with icebergs detaching from glacier termini when the ice connection is no longer able to resist the upward force of flotation and/or the downward force of gravity. The rapid retreat is favored by the thinning of ice near the termini, its flotation, and its weakening by bottom crevasses.

The Columbia tidewater glacier in Alaska retreated around 13 km between 1982 and 2000. Equally the Mendenhall Glacier, which calves into a proglacial lake, has displayed rapid rates of retreat, with 3 km of terminus retreat in the twentieth century (Motyka et al., 2002).

Certainly, calving permits much larger volumes of ice to be lost from the glacier than would be possible through surface ablation (van der Ween, 2002) and great phases of calving in the past help to explain the **Heinrich events** of the Pleistocene and the rapid demise of the Northern Hemisphere ice sheets at the end of the last glacial period.

Some glaciers at high-altitude locations in the tropics have also displayed fast rates of decay over the past 100 or so years, to the extent that some of them now only have around one-sixth to one-third of the area they had at the end of the nineteenth century (Table 11.2) (Kaser, 1999). At present rates of retreat the glaciers and ice caps of Mount Kilimanjaro in East Africa will have disappeared by 2010 to 2020. In the tropical Andes, the Quelccaya ice cap in Peru also has an accelerating and drastic loss over the past 40 years (Thompson, 2000). In neighboring Bolivia, the Glacier Chacaltya lost no less than 40% of its average thickness, two-thirds of its volume and more than 40% of its surface area between 1992 and 1998. Complete extinction is expected within 10 to 15 years (Ramirez et al., 2001).

Table 11.2 Area of tropical glaciers. Source: from data in Kaser (1999, table 2)

Location	*Date*	*Area (km^2)*	*Date*	*Area (km^2)*	*%*
Irian Jaya, Indonesia	*c.* 1850	19.3	*c.* 1990	3.0	15.5
Mount Kenya, Kenya	1899	1.563	1993	0.413	26.4
Lewis Glacier, Kenya	1899	0.63	1993	0.20	31.7
Kibo (Kilimanjaro), Tanzania	*c.* 1850	20.0	1989	3.3	16.5
Ruwenzori, Uganda	1906	6.509	*c.* 1990	1.674	25.7

The reasons for the retreat of the East African glaciers include not only an increase in temperature, but also important may have been a relatively dry phase since the end of the nineteenth century (which led to less accumulation of snow) and a reduction in cloud cover (Mölg et al., 2003).

The fast retreat of glaciers can have a series of adverse geomorphologic effects. Lakes in areas such as the Himalayas of Nepal and Bhutan are rapidly expanding as they are fed by increasing amounts of meltwater. Glacial lake outburst floods are extremely hazardous. Likewise, as slopes are deprived of the buttressing effects of glaciers they can become unstable, generating a risk of increased landsliding and debris avalanches (Kirkbride and Warren, 1999). Increased rates of melting may, for a period of years, cause an increased incidence of summer meltwater floods, but when the glaciers have disappeared, river flow volumes may be drastically reduced (Braun et al., 2000).

Predicted rates of glacier retreat

The monsoonal temperate glaciers of China, which occur in the southeastern part of the Tibetan Plateau, are an example of glaciers that will suffer great

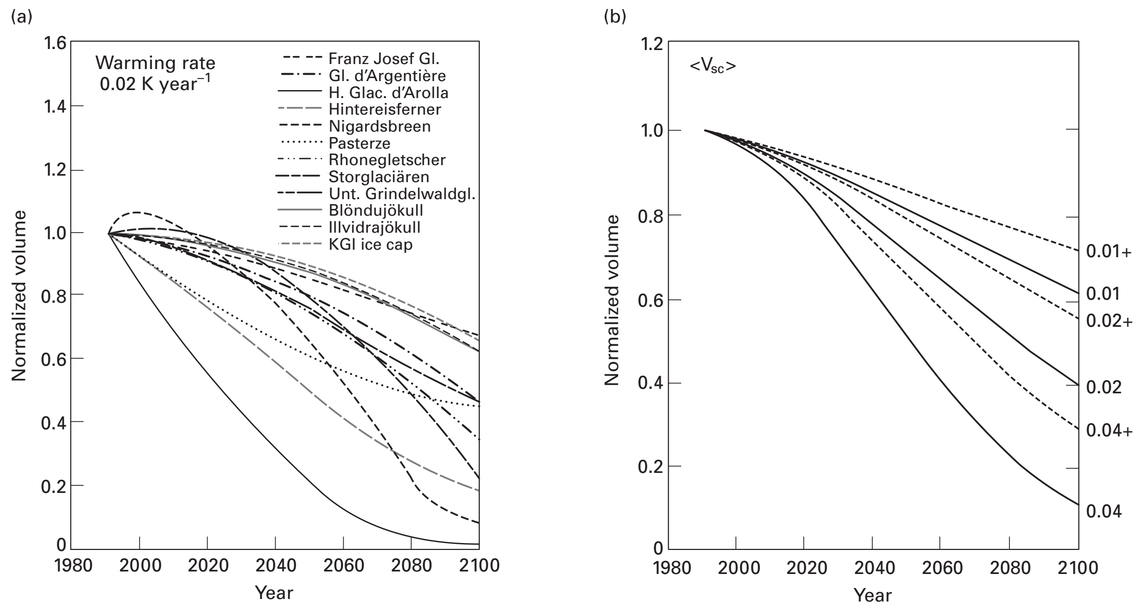

Figure 11.7 Changes in glacier volume 1980–2100 as modeled by Oerlemans et al. (1998) for 12 glaciers: Franz Josef Glacier (New Zealand), Glacier d'Argentière (France), Haut Glacier d'Arolla (Switzerland), Hintereisferner (Austria), Nigardsbreen (Norway), Pasterze (Austria), Rhonegletscher (Switzerland), Storglaciären (Sweden), Unt. Grindelwaldgl. (Switzerland), Blondujökull (Iceland), Illvidrajökull (Iceland), ice cap (Antarctica). (a) Ice-volume change with a warming rate of 0.02 K year^{-1} without a change in precipitation. Volume is normalized with the 1990 volume. (b) Scaled ice-volume change for six climate change scenarios: + refers to an increase in precipitation of 10% per degree warming.

shrinkage if temperatures rise. Indeed, Su and Shi (2002) have, using modeling studies, calculated that if temperatures rise by 2.1°C, these glaciers will lose about 9900 km^2 of their total area of 13,203 km^2 (*c.* 75%) by the year 2100. The subpolar and polar glaciers of China will shrink by *c.* 20% over the same period (Shi and Liu, 2000).

Oerlemans et al. (1998) attempted to model the responses of 12 glaciers (from Europe, New Zealand, and Antarctica) to three warming rates (0.01, 0.02, and 0.04 K per year), with and without concurrent changes in precipitation (Figure 11.7). For a warming rate of 0.04 K per year, without an increase in precipitation, they believed that few glaciers would survive until 2100. On the other hand, with a low rate of warming (0.01 K per year) and an increase in precipitation of 10% per degree of warming, they predicted that overall loss of volume by 2100 would be 10 to 20% of the 1990 volume. On a global basis, mass balance modeling by Braithwate and Zhang (1999) suggested that the most sensitive glaciers in terms of temperature change would be maritime and tropical glaciers. The latter, as in East Africa, are literally hanging on to high peaks, and so are very sensitive to quite modest changes in the height of the snowline.

Sea ice

Since the 1970s, sea ice in the Arctic has declined in area by about 30% per decade (Stocker, 2001: 446). Its thickness may also have become markedly less, with a near 40% decrease in its summer minimum thickness, although the data are imperfect (Holloway and Sou, 2002). By contrast, changes in Antarctic sea ice have been described by the IPCC as 'insignificant' (Stocker, 2001: 446). It has proved difficult to identify long-term trends because of the limited length of observations and the inherent interannual variability of Antarctic sea-ice extent, but it is possible that there has been some decline since the 1950s (Curran et al., 2003).

Figure 11.8 shows the decrease in area of Northern Hemisphere sea ice since 1970 (Vinnikov et al., 1999).

Model projections tend to show a substantial future decrease of Arctic sea-ice cover (Stocker, 2001: 446). By 2050 there will be a roughly 20% reduction in

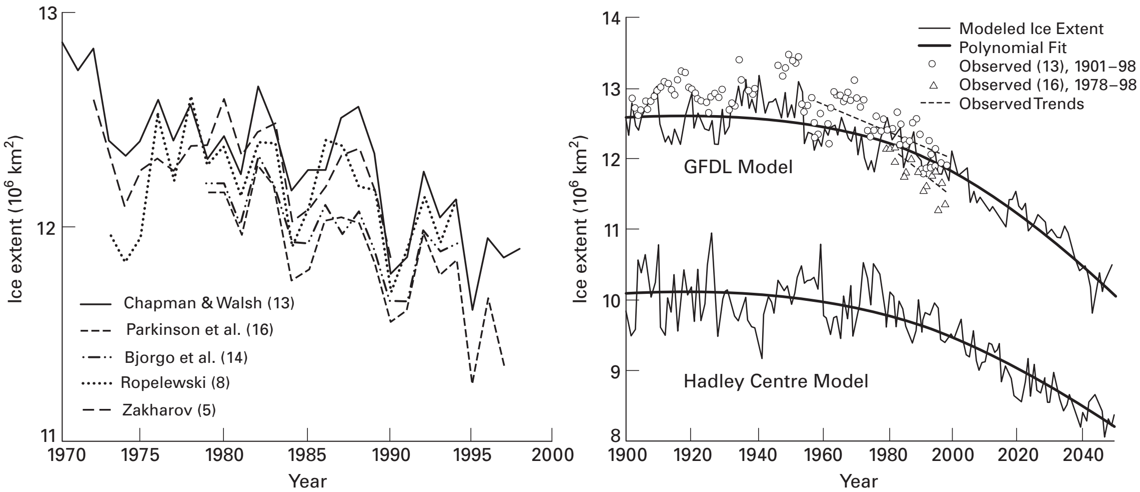

Figure 11.8 Left: observed decrease of Northern Hemisphere sea-ice extent during the past 25 years. Right: observed and modeled variations of annual averages of Northern Hemisphere sea-ice extent. (Source: Vinnikov et al., 1999, figures 1 and 2.)

annual mean Arctic sea-ice extent. A reduction in sea ice area will allow increased absorption of solar radiation so that a further increase in temperature could occur, although this might to some extent be mitigated by increasing cloud cover (Miller and Russell, 2002). However, the IPCC (Anisimov and Fitzharris, 2001: 819) believes that a sea-ice albedo feedback, with amplified warming, means that 'At some point, with prolonged warming a transition to an Arctic Ocean that is ice free in summer – and perhaps even in the winter – could take place.' The IPCC has also recognized that with a doubling of atmospheric greenhouse loadings Antarctic sea ice will undergo a volume reduction of about 25–45%.

Permafrost regions

Regions underlain by permafrost (Figure 11.9) may be especially prone to the effects of global climate change. First of all, the permafrost itself is by its very nature and definition susceptible to the effects of warming. Secondly, the amount of temperature increase predicted for high latitude environments is greater than the global mean, and the loss of permafrost could have a direct effect on atmospheric conditions in the lower atmosphere. Thirdly, permafrost itself is an especially important control of a wide range of geomorphologic processes and phenomena, including slope stability, rates of erosion, ground subsidence, and surface runoff. Fourthly, the nature (e.g., rain rather than snow) and amount of precipitation may also change substantially. Fifthly, the northern limits of some very important vegetation zones, including boreal forest, shrub-tundra and tundra, may shift latitudinally by some hundreds of kilometers. Changes in snow cover and vegetation type may themselves have a considerable impact on the state of permafrost because of their role in insulating the ground surface. For example, if spring temperatures were to increase and early spring snowfall events were to become rain events, the duration of snow cover would decrease, surface albedo would also decrease, leading to an increase in air temperatures, and the snow might provide less insulation. This could cause relatively rapid permafrost degradation (Ling and Zhang, 2003). Conversely, if the warming were to occur in rather more severe periglacial climates, where temperatures are predicted to remain below or at freezing, the winters could become warmer, and wetter, producing an increase in the longevity and depth of the snowpack. This could lead to more insulation (retarding the penetration of the winter cold wave) and a reduction of warming because of an increase in surface albedo. A good general review of the likely consequences of climate change in the tundra of Canada is provided by Smith (1993).

There is some evidence that permafrost has been degraded by the warming of recent decades. For example,

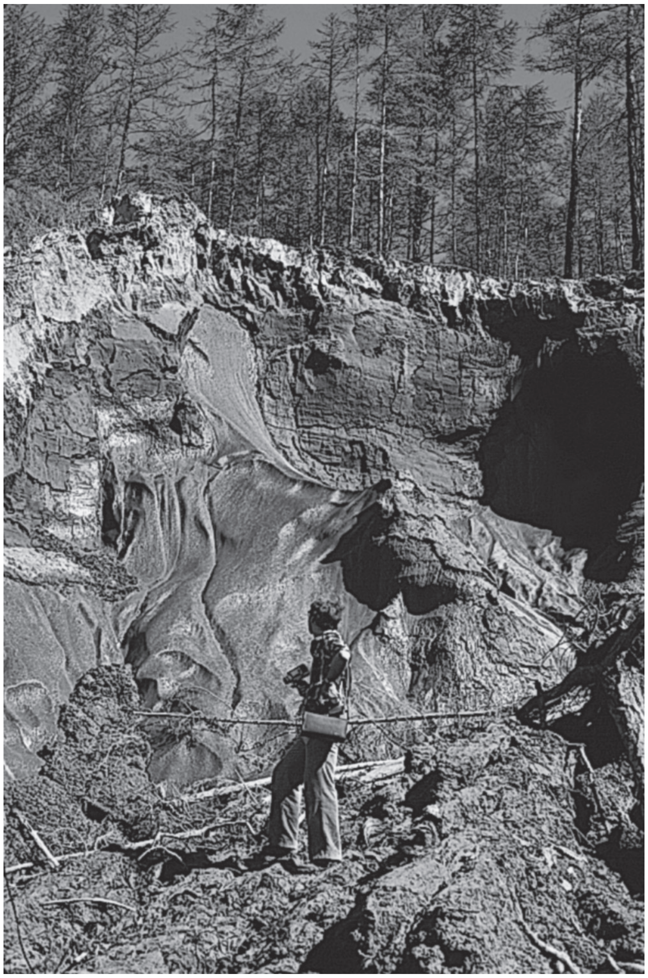

Figure 11.9 Permafrost in Siberia. The great quantity of ice in the permafrost is illustrated behind the figure standing in the foreground (photographed by the late Marjorie Sweeting).

Kwong and Gan (1994) reported a northward migration of permafrost along the Mackenzie Highway in Canada. Between 1962 and 1988 the mean annual temperature in the area rose by 1°C. Over the same period the southern fringe of the discontinuous permafrost zone moved northwards by about 120 km. In Alaska, Osterkamp and Romanovsky (1999) found that in the late 1980s to mid-1990s some areas experienced warming of the permafrost table of 0.5°C to 1.5°C and that associated thawing rates were about 0.1 m per year. However, thawing rates at the base of the permafrost were an order of magnitude slower, indicating (p. 35): 'that timescales of the order of a century are required to thaw the top 10 m of ice rich permafrost, which would be primarily responsible for environmental and engineering problems'.

Recent permafrost degradation has also been identified in the Qinghai–Tibet Plateau (Jin et al., 2000), particularly in marginal areas. Both upward and downward degradation has occurred.

Various attempts have been made to assess future extents of permafrost (Figure 11.10). On a Northern Hemisphere basis, Nelson and Anisimov (1993) have calculated the areas of continuous, discontinuous, and sporadic permafrost for the year 2050 (Table 11.3). They indicate an overall reduction of 16% by that date. In Canada, Woo et al. (1992: 297) suggest that if temperatures rise by 4–5°C, as they may in this high-latitude situation, 'Permafrost in over half of what is now the discontinuous zone could be eliminated'. They add,

> The boundary between continuous and discontinuous permafrost may shift northwards by hundreds of kilometers but because of its links to the tree line, its ultimate position and the time taken to reach it are more speculative. It's possible that a warmer climate could ultimately eliminate continuous permafrost from the whole of the mainland of North America, restricting its presence only to the Arctic Archipelago.

Barry (1985) estimated that an average northward displacement of the southern permafrost boundary by 150 ± 50 km would be expected for each 1°C warming so that a total *maximum* displacement of between 1000 and 2000 km is possible.

Jin et al. (2000: 397) have attempted to model the response of permafrost to different degrees of temperature rise in the Tibetan Plateau region (Table 11.4) and suggest that with a temperature rise of 2.91°C by the end of the twenty-first century the permafrost area will decrease by 58%.

Degradation of permafrost may lead to an increasing scale and frequency of slope failures. Thawing reduces the strength of both ice-rich sediments and frozen jointed bedrock (Davis et al., 2001). Ice-rich soils undergo thaw consolidation during melting, with resulting elevated pore-water pressures, so that formerly sediment-mantled slopes may become unstable. Equally, bedrock slopes may be destabilized if warming reduces the strength of ice-bonded open joints or leads to groundwater movements that cause pore pressures to rise (Harris et al., 2001). Increases in the

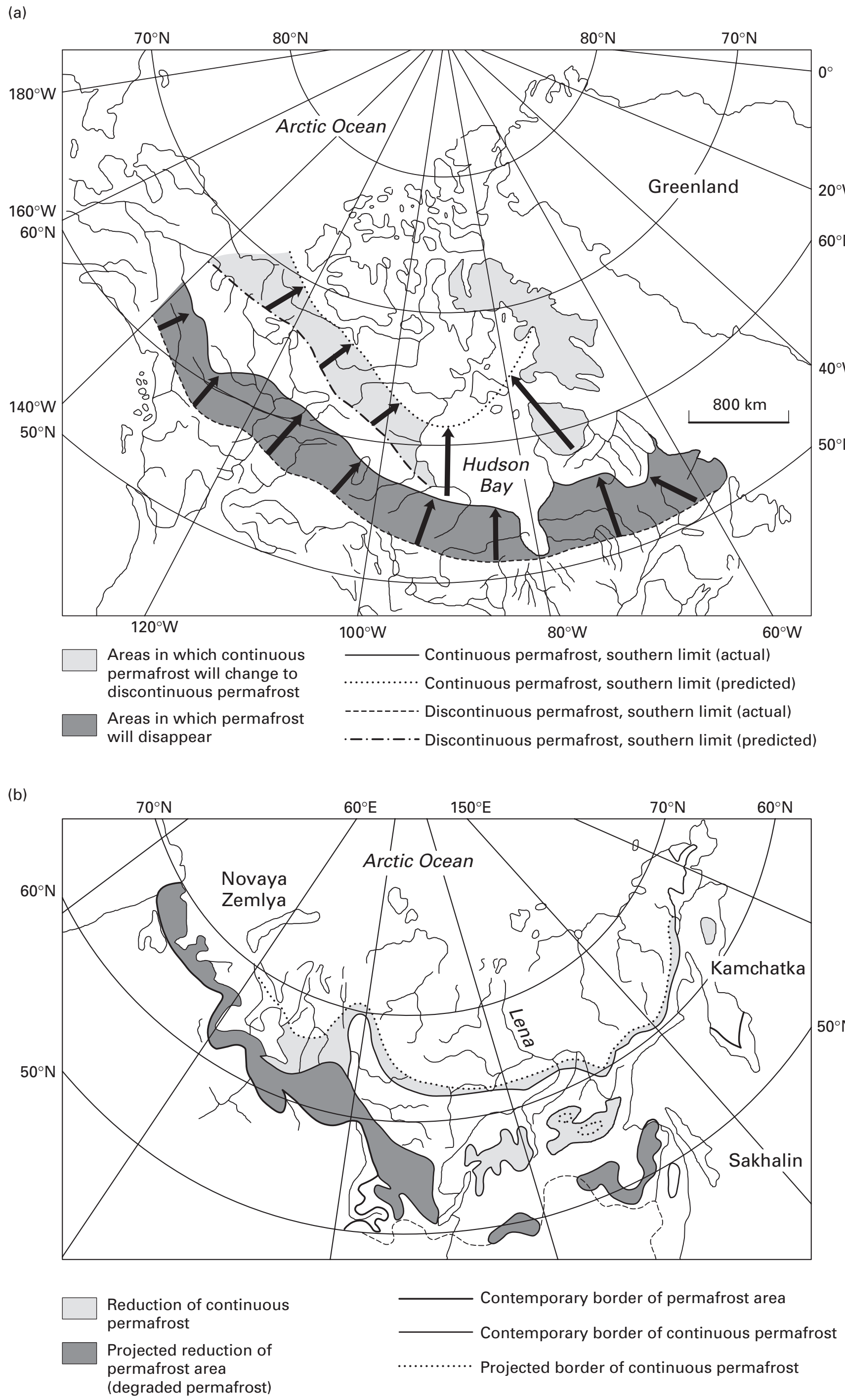

Figure 11.10 Projection of changes in permafrost with global warming for (a) North America and (b) Siberia (after French, 1996, figure 17.5 and Anisimov, 1989).

Table 11.3 Calculated contemporary and future area of permafrost in the Northern Hemisphere by 2050. Source: from Nelson and Anisimov (1993)

Zone	*Contemporary*	*2050*	*Percentage change*
Continuous	11.7	8.5	–27
Discontinuous	5.6	5.0	–11
Sporadic	8.1	7.9	–2
Total	25.4	21.4	–16

Table 11.4 Forecast permafrost degradation on the Qinghai–Tibet Plateau during the next 100 years. Source: Jin et al. (2000, table 11)

Temperature increase (°C) (year)	*Permafrost area (10^6 km^2)*	*Percentage decrease*
0.51 (2009)	1.190	8
1.10 (2049)	1.055	18
2.91 (2099)	0.541	58

thickness of the active layer may make more material available for debris flows.

Likewise, coastal bluffs may be subject to increased rates of erosion if they suffer from thermal erosion caused by permafrost decay. This would be accelerated still further if sea ice were to be less prevalent, for sea ice can protect coasts from wave erosion and debris removal (Carter, 1987). Local coastal losses to erosion of as much as 40 m per year have been observed in some locations in both Siberia and Canada in recent years, while erosive losses of up to 600 m over the past few decades have occurred in Alaska (Parson et al., 2001).

Slope failures in cold environments could be exacerbated by glacier recession (Haeberli and Burn, 2002). The association between glacier retreat since the Little Ice Age and slope movement processes such as gravitational rock slope deformation, rock avalanches, debris flows, and debris slides is evident. This is the case where recent ice retreat has removed buttress support to glacially undercut and oversteepened slopes (Holm et al., 2004).

One of the severest consequences of permafrost degradation is ground subsidence and the formation of thermokarst phenomena (Nelson et al., 2001, 2002). This is likely to be a particular problem in the zones of relatively warm permafrost (the discontinuous and sporadic zones), and where the permafrost is rich in ice (Woo et al., 1992).

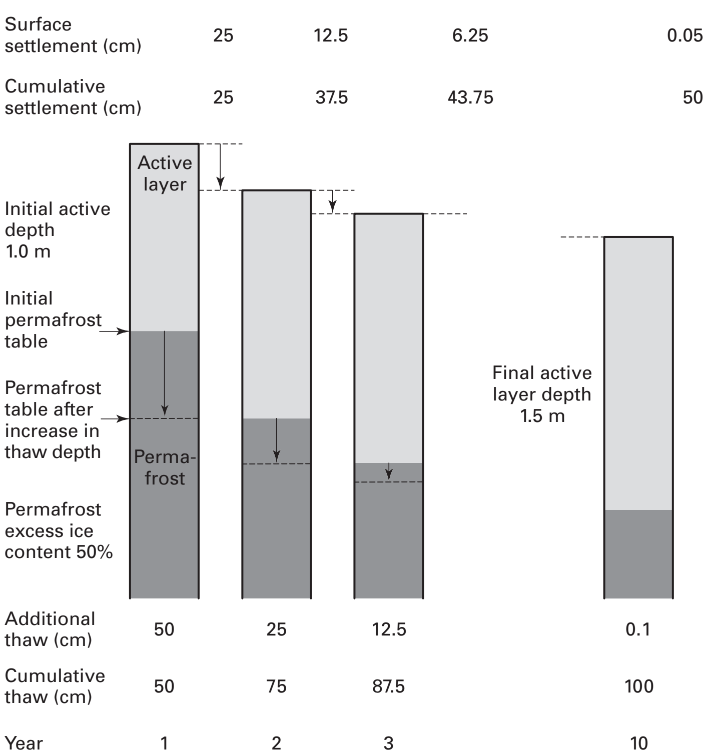

Figure 11.11 Ground settlement in response to a thickening of the active layer in permafrost with an excess ice content of 50%. The active layer increases from 1.0 to 1.5 m and in so doing 1.0 m of permafrost is thawed and the surface settles by 0.5 m. (Source: Woo et al., 1992, figure 8.)

The excess ice content determines how much surface settlement will occur as a result of the thawing of permafrost. It is the volume of water that exceeds the space available within the pores when the soil or sediment is thawed (Woo et al., 1992: 297):

$$EX = 100 \times (V_W / V_T)$$

where EX is the excess ice content (%), V_W is the volume of water that is released on thawing the sample (m^3), and V_T is the total volume of the thawed sample (m^3). Thus, to deepen an active layer by 0.5 m, 1.0 m of permafrost with an EX of 50% must thaw and this ultimately will result in surface settlement of 0.5 m (see Figure 11.11).

Thawing ice-rich areas, such as ice-covered pingos, will settle more than those with lesser ice contents, producing irregular hummocks and depressions. Water released from ice melt may accumulate in such depressions. The depressions may then enlarge into thaw lakes as a result of thermal erosion (Harris, 2002), which occurs because summer heat is transmitted efficiently through the water body into surrounding ice-rich material (Yoshikawa and Hinzman, 2003). Once they have started, thaw lakes can continue to enlarge for decades to centuries because of wave action and continued thermal erosion of the banks. However, if thawing eventually penetrates the permafrost, drainage occurs that leads to ponds drying up.

Increasing temperatures are likely to cause the thickness of the seasonally thawed active layer to become deeper. A simulation by Stendel and Christensen (2002) indicated a 30–40% increase by the end of the twenty-first century for most of the permafrost area in the Northern Hemisphere.

Points for review

What are the main components of the cryosphere?
How are valley glaciers likely to respond to global warming?
Will the polar ice sheets respond catastrophically to global warming?
What are the consequences of permafrost melting?

Guide to reading

French, H. M., 1996, *The periglacial environment* (2nd edn). Harlow: Longman. A general text on periglacial areas, but with a summary of the effects of global change.

Harris, C. and Stonehouse, B. (eds), 1991, *Antarctica and climatic change*. London: Belhaven Press. A study of the response of Antarctica to future warming and a consideration of the global implications.

Wadhams, P., Dowdeswell, J. A. and Schofield, A. N. (eds), 1996, *The Arctic and environmental change*. Amsterdam: Gordon and Breach. A collection of papers presented to the Royal Society of London and covering many aspects of cryospheric change in northern regions.