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The Cryosphere: Changes and Their Impacts

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CONTENTS

EXECUTIVE SUMMARY

The cryosphere covers a substantial amount of the Earth's surface and is very sensitive to climate change. This chapter updates the 1990 IPCC Impacts Assessment—which examined snow, glaciers, and permafrost—but also considers snow avalanches and sea, river, and lake ice. Impacts on feedbacks to global climate and sea level caused by changes in the cryosphere are covered in other chapters of the IPCC Second Assessment Report.

This chapter arrives at the following conclusions:

- Many components of the cryosphere are sensitive to changes in atmospheric temperature because of their thermal proximity to melting. The extent of glaciers has often been used as an indicator of past global temperatures (High Confidence).
- Projected warming of the climate will reduce the area and volume of the cryosphere. This reduction will have significant impacts on related ecosystems, associated people, and their livelihoods (High Confidence).
- There will be striking changes in the landscapes of many high mountain ranges and of lands at northern high latitudes (High Confidence). These changes may be exacerbated where they are accompanied by growing numbers of people and increased economic activity (Medium Confidence).

From an examination of past measurements of the cryosphere, the following 20th-century trends are observed:

- Obvious thinning, mass-loss, and retreat of mountain glaciers (High Confidence): The extent of alpine ice in the European Alps probably is more reduced today than at any time during the past 5,000 years.
- Borehole measurements show that permafrost is warming in some areas but not everywhere (Medium Confidence).
- Later freeze-up and earlier break-up dates for river and lake ice in the tundra and boreal lands: These are at least a week different compared to last century (Medium Confidence).
- No convincing evidence of trends in Antarctic or Arctic sea-ice extent (Low Confidence).
- Much variability of seasonal snow from year to year but no definitive trends, except that the areal extent of Northern Hemisphere continental snow cover has decreased since 1987 (High Confidence).
- Little change in the gross features of ice sheets (Medium Confidence).

If projections of climate for the year 2050 are realized (UKMO transient experiment data; Greco *et al*., 1994), then the following impacts on the cryosphere are likely:

- Pronounced reductions in seasonal snow, permafrost, glacier, and periglacial belts of the world, with a corresponding shift in landscape processes (High Confidence).
- Disappearance of up to a quarter of the presently existing mountain glacier mass (Medium Confidence).
- Increases in the thickness of the active layer of permafrost and the disappearance of extensive areas of discontinuous permafrost in continental and mountain areas (High Confidence).
- Less ice on rivers and lakes. Freeze-up dates will be delayed, and break-up will begin earlier. The river-ice season could be shortened by up to a month. Many rivers within the temperate regions will become ice-free or develop only intermittent or partial ice coverage (Medium Confidence).
- A large change in the extent and thickness of sea ice, not only from warming but also from changes in circulation patterns of both atmosphere and oceans. There is likely to be substantially less sea ice in the Arctic Ocean (Medium Confidence).
- Major changes in the volume and extent of ice sheets and deep, continuous permafrost are unlikely by 2050 because they are very cold and react with longer time lags (High Confidence); however, unforeseen changes in the West Antarctic ice sheet still could occur (Low Confidence).

As a result of these changes in the cryosphere, the following impacts on other systems are expected:

- More water will be released from regions with extensive glaciers (High Confidence). In some semi-arid places near high mountains, such as in central Asia and Argentina, this glacial runoff may increase water resources. In other places, summer water resources may diminish as glaciers disappear (Medium Confidence).
- In temperate mountain regions, reduced snow cover will cause moderation of the seasonal flow regime of rivers, so that winter runoff increases and spring runoff decreases (Medium Confidence). Such changes may benefit the hydroelectricity industry (Medium Confidence).
- Reduced snow cover and glaciers will detract from the scenic appeal of many alpine landscapes

(Medium Confidence). For temperate mountains, less snow will restrict alpine tourism and limit the ski industry to higher alpine areas than at present. Snow seasons will tend to be shorter and less reliable, and there will be detrimental socioeconomic impacts on mountain communities that depend on winter tourism.

- Widespread loss of permafrost over extensive continental and mountain areas will trigger erosion or subsidence of ice-rich landscapes, change hydrologic processes, and release carbon dioxide and methane to the atmosphere (Medium Confidence).
- Cryospheric change will reduce slope stability and increase the incidence of natural hazards for people, structures, and communication links in mountain lands and continental permafrost areas. Buildings, other structures, pipelines, and communication links will be threatened (Medium Confidence).
- Engineering and agricultural practices will need to adjust to changes in snow, ice, and permafrost distributions (High Confidence).
- Thawing of permafrost could lead to disruption of existing petroleum production and distribution systems in the tundra, unless mitigation techniques are adopted. Reduced sea ice may aid new exploration and production of oil in the Arctic basin (Medium Confidence).
- Less sea ice could reduce the renewal of deep waters of the North Atlantic, affect the ocean conveyor system, decrease albedo, and consequently induce climate feedbacks (Low Confidence).
- Improved opportunities for water transport, tourism, and trade at high latitudes are expected from a reduction in sea, river, and lake ice. These will have important implications for the people and economies of the Arctic rim (Medium Confidence).

7.1. Introduction

The cryosphere, which represents all global snow, ice, and permafrost, contains nearly 80% of all freshwater. It includes seasonal snow, mountain glaciers, ice caps, ice sheets, seasonally frozen soils, permafrost, river ice, lake ice, and sea ice (Table 7-1). Permafrost underlies as much as 25% of the global land surface. Seasonal snow has the largest area of any component of the cryosphere; at its maximum in late winter it covers almost 50% of the land surface of the Northern Hemisphere. A huge proportion of the mass of the cryosphere is contained in ice sheets, but at time scales of a century or less they are least sensitive to climate change.

The cryosphere is an important part of Earth's geographical and climate systems, and its components change over diverse time scales. Areas of snow and sea ice expand and contract markedly with the seasons (Table 7-1), and ice sheets underwent vast change as the Earth became warmer or cooler from the last interglacial period. During the previous interglacial—when global temperature was higher than at present—the global cryosphere was smaller, but it had twice its current mass during the last glaciation.

This chapter brings up to date the 1990 IPCC Impacts Assessment that dealt with seasonal snow cover, ice, and permafrost (Street and Melnikov, 1990) and the IPCC 1992 Supplementary Report (Melnikov and Street, 1992). Snow avalanches and sea, river, and lake ice are considered for the first time, and substantial new material has been added on glaciers and permafrost. How each component of the cryosphere is changing and its sensitivity to climate change is assessed, as is the impact of climate change on the cryosphere by about 2050. How these changes in the cryosphere might affect other physical and human systems is also described. Critical information we need to know is identified in the final section.

Table 7-1: *Estimate of the size of the cryosphere (modified from U.S. DOE, 1985; Barry, 1985; Street and Melnikov, 1990; Meier, 1993; Gloersen et al., 1992; and Chapter 7, Changes in Sea Level, of the Working Group I volume).*

Source	Area (10^6 km^2)	Ice Volume (10^6 km^3)
Seasonal Snow		
N. Hemisphere winter	46.3	<0.01
N. Hemisphere summer	3.7	
S. Hemisphere winter	0.9	
S. Hemisphere summer	<0.1	
Ice Caps and Glaciers	0.6	0.09
Ice Sheets		
Greenland	1.7	2.95
West Antarctica	2.4	3.40
East Antarctica	9.9	25.92
Antarctic ice shelves	1.6	0.79
Permafrost	25.4	0.16
River and Lake Ice	<1.0	
Sea Ice		
N. Hemisphere winter	16.0	0.05
N. Hemisphere summer	9.0	0.03
S. Hemisphere winter	19.0	0.03
S. Hemisphere summer	3.5	<0.01

Although the cryosphere forms an integral part of the climate system, the ways in which it generates important feedbacks into the climate system are not discussed because this information is contained in the Working Group I volume. Similarly, the interaction of ice sheets and sea-level change are considered in detail in the Working Group I volume (see Chapter 7, *Changes in Sea Level*). A summary of ice sheets and climate change is offered here in Box 7-1.

7.2. Is the Cryosphere Changing?

7.2.1. Snow

Snow tends to be a very transient part of the Earth's surface, often lasting for only a few days or months. Monitoring of global seasonal snow is practical only with satellite remote sensing, so there are no reliable records prior to 1971. Records since then show considerable variability in the Northern Hemisphere continental snow cover from year to year (Figure 7-1), which makes long-term trends difficult to detect. The extent of snow has been less since 1987 (Robinson *et al.,* 1993), with the largest negative snow anomalies occurring in spring. Recent analysis of directly measured snow cover over the North American Great Plains (Brown *et al.,* 1994) reveals

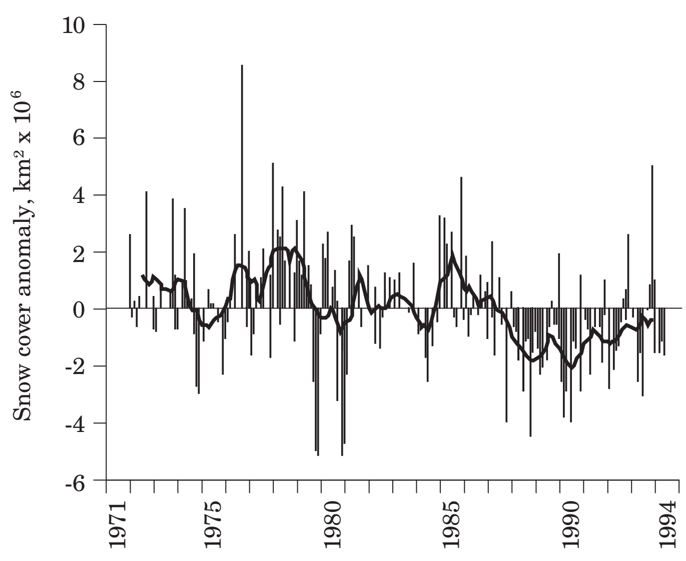

Figure 7-1: Seasonal snow cover anomalies in the Northern Hemisphere for 1971–1994 (after Robinson *et al.*, 1993). Each vertical line represents a season, with the dark line a 12-month running mean.

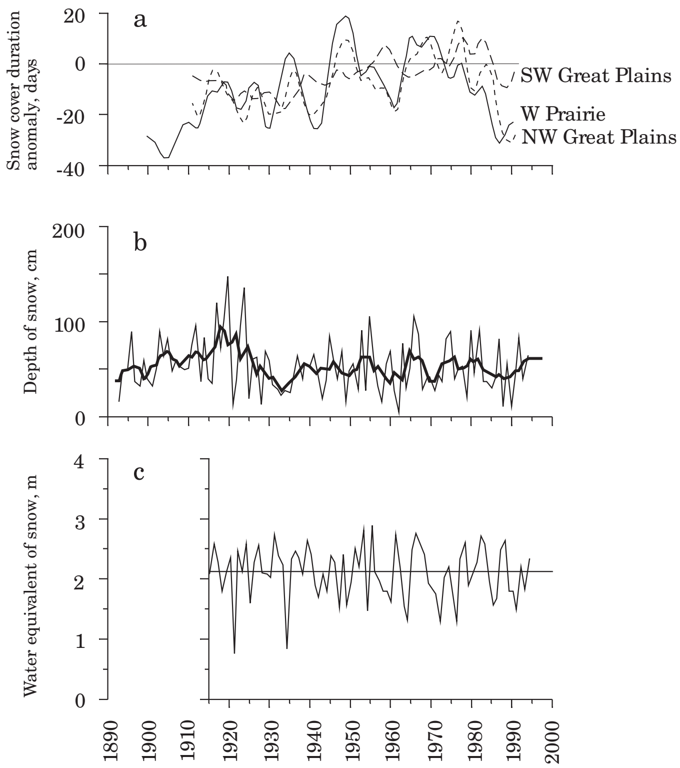

Figure 7-2: Historical variability of snow for three different settings: (a) Snow-cover duration anomalies over the continental interior of North America (after Brown *et al*., 1994); (b) depth of seasonal snow at an alpine area at Davos, elevation 1560 m, in Switzerland, as measured every 1 January 1893–1994 (updated from Foehn, 1990); and (c) water equivalent of snow near the equilibrium line of a glacier at Claridenfirn, elevation 2700 m, in Switzerland, as measured every spring 1914–1993 (updated from Müller and Kaeppenburger, 1991).

an increase over the past century (Figure 7-2a) but a decline over the Canadian Prairies.

The longest alpine snow-cover time series come from the European Alps. Recently, there has been a trend to lower snow depths in early winter but delayed ablation in spring; however, the century-long record for Davos (Figure 7-2b) shows no obvious trend in snow depth but large interannual variability (Foehn, 1990). This lack of any trend is reinforced by Figure 7-2c, which shows snow water equivalent near the glacier equilibrium line at Claridenfirn (Müller and Kaeppenberger, 1991). Elsewhere, a remarkable decrease in accumulated and maximum snow depth is reported since the 1987–88 winter over a wide area of western Japan. Here, snowfall normally is heavy and is affected by the winter monsoon (Morinaga and Yasunari, 1993).

Few observations assess long-term trends in the structure of snow. In the European Alps, it is summarized by ram (ramsonde) profiles, which are catagorized into six main types. Each type has a certain potential for snow avalanche formation and is linked to a specific winter climate. Over the last 50 years in Switzerland there has been no trend in ram profiles. A few long-term records of avalanche occurrence have been assembled, notably from Iceland (Bjornson, 1980), Europe (Fitzharris and Bakkehoi, 1986), western Canada (Fitzharris and Schaerer, 1980), and the United States (Armstrong, 1978). They demonstrate that although avalanches reach catastrophic proportions in a few winters, there are no clear temporal trends or regular periodicities.

7.2.2. Ice Caps and Glaciers

Internationally coordinated, long-term monitoring of glaciers started in 1894 and today involves collection and publication of standardized information on the distribution and variability of glaciers over space (glacier inventories) and time (glacier fluctuations: mass balance, length change). Information on special events (instabilities, catastrophic changes) also is available (UNEP, 1993). The results of long observational series on fluctuations of mountain glaciers represent convincing evidence of past climatic change on a global scale [e.g., IAHS(ICSI)/UNEP/UNESCO, 1988, 1989, 1993, 1994].

Mass loss and retreat of glaciers is common in many mountain areas of the world. Oerlemans (1994) looks at glacial retreat on a global level, using a scaling system to classify different glaciers. He concludes that during the period 1884–1978, mean global glacial retreat corresponded to a calculated warming of 0.66 ± 0.10°C per century. Chapter 7*, Changes in Sea Level*, of the Working Group I volume suggests that glaciers have lost sufficient ice over the last 100 years to raise sea level by 0.2–0.4 mm/yr. Recent analyses by Oerlemans and Fortuin (1992), Meier (1993), and Dyurgerov (1994) all show negative global ice-mass balances during this century. Although the glacial signal appears homogeneous at the global scale, there is great variability at local and regional scales and over shorter time periods of years to decades (Letréguilly and Reynaud, 1989).

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. On average, this rate of icemelt is roughly one order of magnitude higher than the overall mean calculated for the end of the last glaciation and is broadly consistent with anthropogenic greenhouse forcing of 2–3 W/m^2. The recent discovery of a stone-age man from cold ice/permafrost on a high-altitude ridge of the Oetztal Alps confirms the results of earlier moraine investigations: The extent of Alpine ice probably is more reduced today than at any time during the past 5,000 years. Glacier wastage in the European Alps appears to be accelerating (Haeberli, 1994).

In the circum-Arctic, there is a consistent tendency for negative mass balances over the past 30 years. Mass balances of two Canadian ice caps are slightly negative (Koerner and Brugman, 1991), but these are so cold that the main signal is likely to be in the change of firn temperatures. The general picture from West Greenland is that of a strong retreat through this century, with a trim line zone around many glacier lobes. For the North Greenland mountain glaciers and ice caps, the situation is less clear, but the general impression is that changes are small. East

Greenland mountain glaciers seem to have behaved as the West Greenland ones, but surging advances occur frequently in certain regions (Weidick, 1991a-d). The longest glacier record for northern Sweden shows a preponderance of negative mass-balance years since 1946 (Letréguilly and Reynaud, 1989). In Spitsbergen, several glaciers have been losing mass (Hagen and Liestol, 1990), although the extent of glaciers probably was less about 5,000 years ago (Fujii *et al.,* 1990).

In Asia, the area of glaciers in Kazakhstan reduced by 14%, their number diminished by 15%, and the general volume of ice reduced by 11% between 1955 and 1979 (Glazyrin *et al.,* 1986, 1990; Dikikh and Dikikh, 1990; Kotlyakov *et al.,* 1991; Popovnin, 1987). Fluctuations of 224 glaciers in Central Asia from the 1950s to the 1980s can be summarized as retreating (73%), advancing (15%), and stable (12%) (Shi and Ren, 1988, 1990).

The general picture of ice retreat continues for mountains in the tropics. It began around the middle of the 19th century in the Ecuadorian Andes and in New Guinea but only after 1880 in East Africa (Hastenrath, 1994). Schubert (1992) documents fast, ongoing glacier shrinkage in Venezuela, as do Hastenrath and Ames (1995) for the Yanamarey glacier in the Cordillera Blanca of Peru.

The Southern Hemisphere record is not as detailed as that for the Northern Hemisphere. In New Zealand, Ruddell (1990) shows that most glaciers have retreated during the 20th century, some by three kilometers or more. The surface of the Tasman Glacier has thinned by more than 100 m. This widespread recession has reversed for western glaciers since about 1983. In South America, the Upsala glacier has retreated about 60 m/yr over the last 60 years, and this rate seems to be accelerating (Malagnino and Strelin, 1992). The area of the South Patagonian Ice Field has diminished by about 500 km^2 from 13,500 km^2 in 41 years. Surface lowering also has been considerable—more than 100 m at ablation areas of some glaciers (Aniya *et al.,* 1992). The Soler Glacier thinned at a rate of 5.2 m/yr from 1983 to 1985 (Aniya *et al.,* 1992), and the surface of Tyndall Glacier (one of the southernmost outlet glaciers) lowered by 20 m between 1985 and 1990 (a rate of 4.0 m/yr), according to Kadota *et al*. (1992). A few glaciers are advancing: The Pio XI glacier in Patagonia is larger now than it has been at any time in the past 6,000 years (Warren, 1994).

Frontal positions of alpine glaciers of the Dry Valleys in Antarctica have fluctuated, with no apparent trends (Chinn, 1993). On sub-Antarctic Heard Island, in the Indian Ocean, there has been widespread retreat since 1947—with some small glaciers decreasing in area by as much as 65% (Allison and Keate, 1986).

7.2.3. Permafrost

Permafrost—ground material that remains below freezing—underlies continental areas in the tundra and some boreal lands to considerable depths. It also is present under shallow polar seabeds, in ice-free areas in Antarctica, on some sub-Antarctic islands, and in many mountain ranges and high plateaus of the world (Cheng and Dramis, 1992; Harris and Giardino, 1994; King and Åkermann, 1994; Qiu, 1994). Figure 7-3a shows the contemporary distribution of permafrost in the Northern Hemisphere. Some is relict, having formed during colder glacial periods, but has survived due to the negative heat balance at the ground surface or the very long time it takes for deep permafrost to thaw. There also are many sedimentary structures that show that permafrost was once more extensive than today.

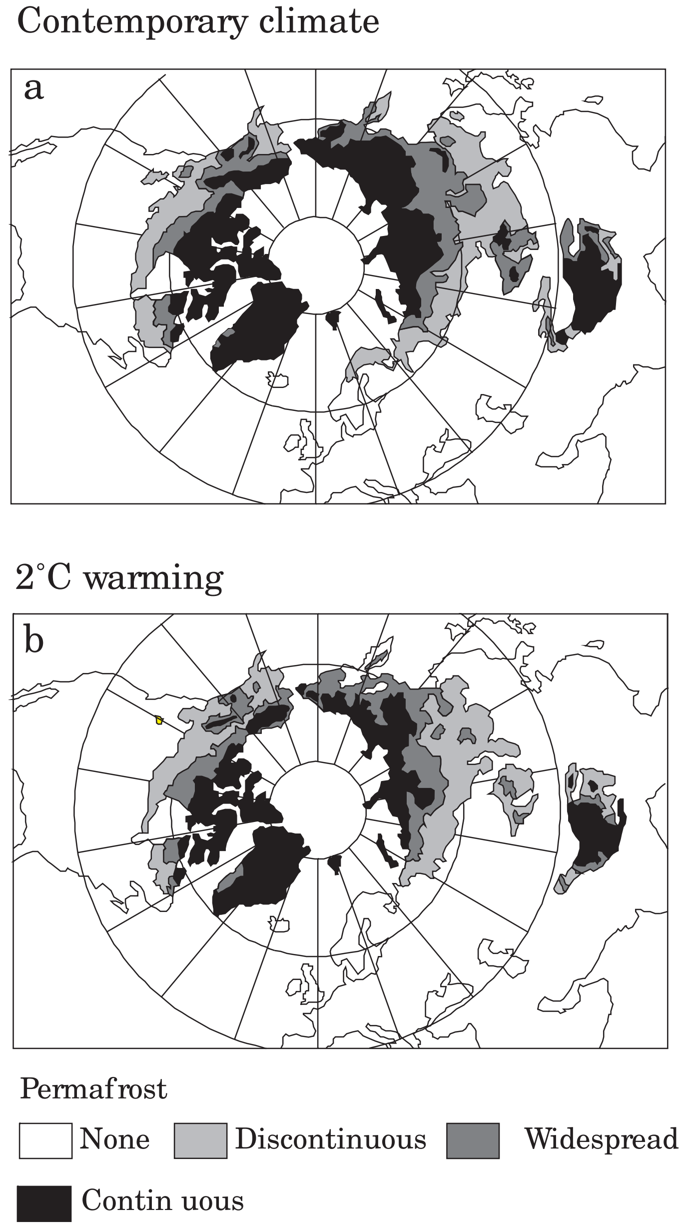

Figure 7-3: Distribution of Northern Hemisphere permafrost for (a) the present and (b) 2050 (based on Nelson and Outcalt, 1987; Anisimov and Nelson, 1995).

Long-term measurements in deep boreholes in Alaska (Harrison, 1991; Zhang and Osterkamp, 1993), Canada

(Taylor, 1991), and elsewhere (Koster *et al.,* 1994) demonstrate a distinct but spatially heterogeneous warming trend in lowland permafrost. In northern Alaska, Lachenbruch and Marshall (1986) have demonstrated a warming of the permafrost of 2–4°C over the last century. Temperatures along the coast have varied over a range of 4°C (Osterkamp, 1994). During the last two decades, permafrost in Russia and China also has warmed (Pavlov, 1994; Wang and French, 1994). Some discontinuous permafrost in the southern half of Alaska is currently thawing (Osterkamp, 1994).

First attempts are now being made to monitor the long-term evolution of high-mountain permafrost by aerial photogrammetry of permafrost creep, borehole measurements of permafrost deformation and temperature, data archiving from geophysical surface soundings for later repetition, and qualitative analysis of infrared aerial photography (Francou and Reynaud, 1992; Haeberli *et al.,* 1993; Harris, 1990; Vonder Mühll and Schmid, 1993; Wagner, 1992).

7.2.4. River and Lake Ice

Chronologies of river and lake ice formation and disappearance provide broad indicators of climate change over extensive lowland areas, in much the same way that glaciers provide indicators for mountains (Palecki and Barry, 1986; Reycraft and Skinner, 1993). River-ice data have been summarized for homogenous hydrologic regions of the former Soviet Union (FSU) over the period 1893–1985 and adjusted to account for any effects of water-resource development (Soldatova, 1992). Although there is appreciable interdecadal variability, there are significant long-term spatial patterns and temporal trends. Freeze-up on rivers such as the Danube, Dnieper, Don, Lower Volga, and rivers of the Black Sea region is now delayed by an average of two to three weeks compared with the early part of the record. Further east, a weaker trend to earlier freeze-up dates is observed for the Yenisey and Lena (Ginzburg *et al.,* 1992).

The same broad-scale pattern is evident for break-up dates (Soldatova, 1993). Break-up on major rivers such as the Upper Volga, Oka, Don, Upper Ob, and Irtysh has advanced by an average of 7–10 days during the last century. In some rivers, such as the Lower Don, the overall result is a reduction in the winter ice season by as much as a month. In Central and Eastern Siberia (e.g., Middle to Lower Yenisey and Upper Lena), some rivers exhibit later break-up dates and, hence, an overall expansion of the ice season. In northern Scandinavia, historical records as far back as 1693 indicate that break-up in the Tornelven river is occurring much earlier during the 20th century than in earlier times (Zachrisson, 1989).

Assel *et al.* (1995) analyze lake-ice freeze-up and break-up in North America using records from 1823–1994 at six sites throughout the Great Lakes. Freeze-up dates gradually become later and ice-loss dates gradually earlier from the beginning of the record to the 1890s but have remained relatively constant during the 20th century. Ice-loss dates at deeper-water environments with mixing of offshore waters were earlier during the 1940s and 1970s but later during the 1960s. Global warming during the 1980s was marked by a trend toward earlier ice-loss dates. Other North American lake-ice chronologies are presented by Hanson *et al.* (1992) and Reycraft and Skinner (1993).

7.2.5. Sea Ice

The most effective means of measuring sea-ice extent is through the use of satellite-borne passive microwave radiometers, but these data are available only from 1973 onward. They show no convincing evidence of trends in global sea-ice extent (Gloersen *et al.,* 1992; Parkinson and Cavalieri, 1989; Gloersen and Campbell, 1991). Studies of regional changes in both the Arctic and Antarctic indicate some trends (e.g., Mysak and Manak, 1989; Gloersen and Campbell, 1991; Parkinson, 1992), but longer data sets are needed because they are only of decadal length.

Knowledge of the regional variability of ice thickness comes almost entirely from upward sonar profiling by submarines (Wadhams, 1990a; Wadhams and Comiso, 1992), such as the large-scale maps of mean ice thickness in the Arctic shown in Figure 7-4 (based on Sanderson, 1988; Vinje, 1989; Wadhams, 1992). Variations in ice thickness are in accord with the predictions of numerical models, which take account of ice dynamics and deformation, as well as ice thermodynamics. Measurements from a series of submarine transects near the North Pole show large interannual variability in ice draft over the period 1979–1990 (McLaren *et al.,* 1992). There is evidence of a decline in mean thickness in the late 1980s relative to the late 1970s (Wadhams, 1994). Etkin (1991) found that ice breakup correlated well with melting degree-days in some areas of Hudson Bay and James Bay (Canada), but in other areas of the bays it was more strongly influenced by ice advection, freshwater inflow, and conditions within the air-water-ice boundary layer.

Winter pack ice around Antarctica has major climatic importance. The annual variation in sea-ice area is very large—from 3 x 10^6 km^2 in February to 19 x 10^6 km^2 in September (Gloersen *et al.,* 1992). No change in the thickness of Antarctic sea ice can be detected from the limited information available. The only systematic data have been obtained by repetitive drilling in a region of first-year ice from the eastern Weddell-Enderby Basin. The modal ice thickness is 0.5–0.6 m, and maximum observed keel drafts are about 6 m. In the limited regions of the Antarctic where multiyear ice occurs, there is a preferred ice thickness of about 1.4 m (Wadhams and Crane, 1991; Lange and Eicken, 1991).

7.3. How Sensitive is the Cryosphere to Climate Change?

7.3.1. Snow

Snow cover in temperate regions is generally thin (a few meters to a few centimeters) and often close to its melting

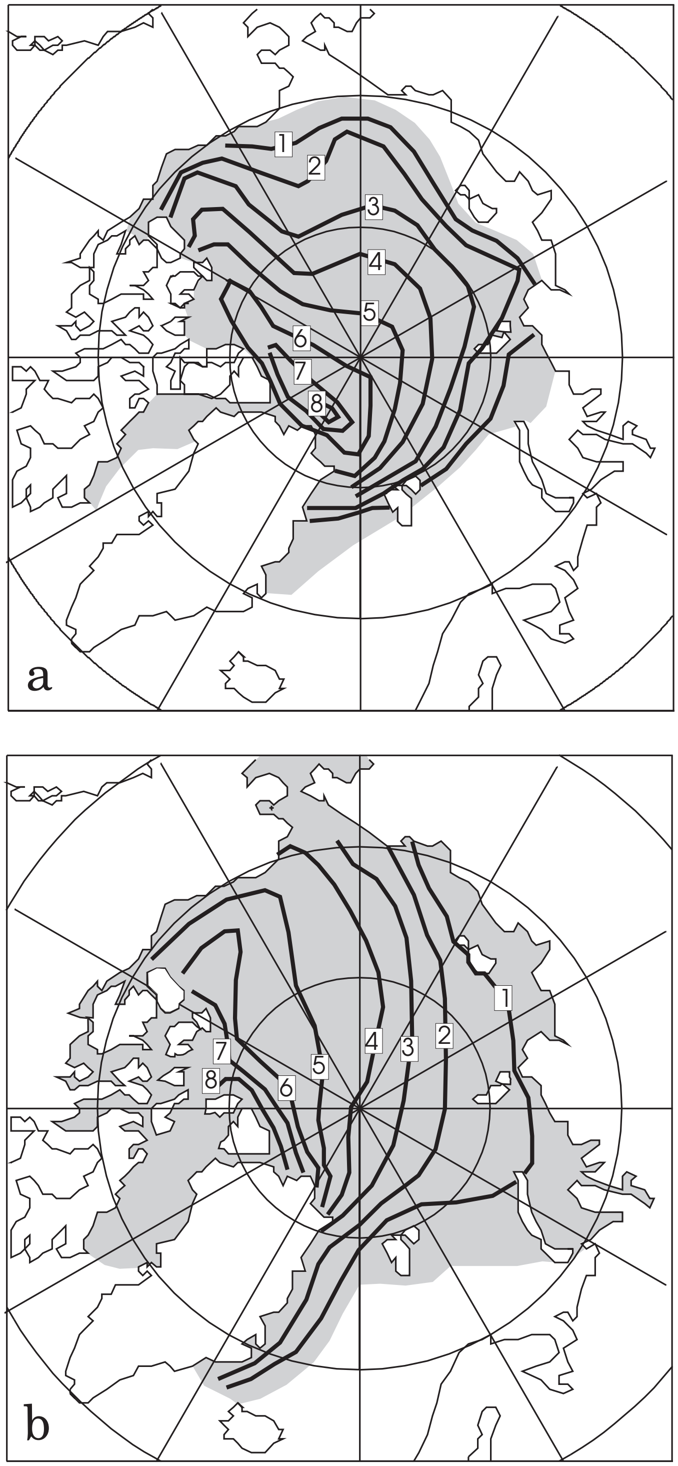

Figure 7-4: Estimated mean thickness of Arctic Basin sea ice in meters for (a) summer and (b) winter (after Sanderson, 1988; Bourke and Garrett, 1987). The data do not include open water, thus overestimating mean ice draft.

point; consequently, both continental and alpine snow covers are very sensitive to climate change. Karl *et al*. (1993) found that a 1°C increase in the annual temperature of the Northern Hemisphere results in a 20% reduction in North American snow cover. Snow accumulation and melt models can estimate water stored as seasonal snow in alpine areas and can be used to examine its sensitivity to climate change. For the Southern Alps of New Zealand, Fitzharris and Garr (1995) found that water stored as seasonal snow declines as temperature increases and precipitation decreases but that the relationship is not linear. Seasonal snow is more sensitive to decreases in temperature than to increases. As precipitation increases, snow accumulation becomes less sensitive to temperature changes. The volume of water stored also depends on catchment hypsometry. In Australia, the duration of snow cover is found to be very sensitive to changes in temperature. Large increases in precipitation (50%) are necessary to offset even a 0.5°C warming (CSIRO, 1994).

Large, catastrophic avalanches are mostly the result of special weather situations lasting for 5 to 10 days. In the Swiss Alps, only nine weather types out of a possible twenty-nine have been responsible for past catastrophic avalanche cycles. These large avalanche episodes are thus sensitive to the frequency of particular weather types. Similar findings also come from Norway (Fitzharris and Bakkehoi, 1986) and Canada (Fitzharris, 1987).

7.3.2. Ice Caps and Glaciers

In the chain of processes linking climate and glacial fluctuations, mass balance is the direct, undelayed reaction; glacier length variation is the indirect, delayed response. Averaged for glaciers with comparable geometry and over time intervals of decades, changes in length provide an integrated and smoothed signal of climate change. At shorter time scales, glaciers will vary markedly with climate change, especially temperature. The nature of the response to warming will vary from glacier to glacier depending on accompanying precipitation change, the mass-balance gradient, and hypsometry.

Most glaciers of the world are more sensitive to changes in temperature than to any other climatic element. In the case of many Asian glaciers, where precipitation occurs mainly during the summer monsoon season, temperature has a double impact. The first impact is an increase in the absorption of solar radiation due to a lowering of the surface albedo as snowfall is converted to rainfall. The second effect is an increase in the energy exchange between the atmosphere and the glacier surface (Ageta and Kadota, 1992). Many maritime glaciers, which have large mass turnover, are more sensitive to changes in precipitation than to temperature. This complex dependence of glacier mass balance on temperature, precipitation, and radiation makes it difficult to define their sensitivity to climate change.

A number of approaches are used to express the sensitivity of glaciers to climate. Where there are long glacier records for calibration, regression equations are used to relate mass balance to summer mean temperature and annual (or winter) precipitation (Laumann and Tvede, 1989; Chen, 1991). Changes in summer ablation rates at the margins of Greenland glaciers are nearly linear with changes in summer temperature, with sensitivities ranging from 0.43–0.57 m/yr/°C (water equivalent; Braithwaite and Olesen, 1990). Although these empirical relationships are valuable for regions in which they have been developed, they are not always applicable to other mountain areas with different climate and terrain. The sensitivity of glacier ablation has been assessed for number of melting degree-days (dd) for a range of glaciers by Braithwaite and Olesen (1989, 1990, 1993). Ablation sensitivity for ice varies from 5.5 to 7.6 mm/dd (Braithwaite and Oleson, 1989). Values for snow range from 1.0 to 5.7 mm/dd (Johannesson *et al.,* 1993). Hydrometeorological models are similar but include precipitation as well as temperature; they also usually estimate the mass balance (e.g., Woo and Fitzharris, 1992) and sometimes the water balance (e.g., Tangborn, 1980). Energy-balance models also have been used to assess the sensitivity of glacier ablation to climate (Oerlemans, 1991; Oerlemans *et al.,* 1993); they are appealing because they are more physically based, but they require input parameters that are difficult to obtain.

Glaciers are sensitive to changes in atmospheric circulation patterns, but it is difficult to quantify the response. The behavior of the Franz Josef glacier in New Zealand this century can be qualitatively explained by changes in the westerlies and shifts in the subtropical high-pressure zone (Fitzharris *et al.,* 1992). A major advance since 1983 appears to be consistent with circulation changes induced by three large El Niño events. In the first half of the 20th century, warming accounted for most of the ice thinning observed on Mount Kenya. However, from 1963 to 1987, greater atmospheric humidity was instrumental in enhancing melt—perhaps a consequence of enhanced evaporation in the Indian Ocean and its subsequent advection over East Africa by the prevailing atmospheric circulation patterns (Hastenrath, 1992, 1994; Hastenrath and Kruss, 1992). In the Canadian High Arctic, mass-balance and ice-core melt studies show that when the circumpolar vortex shifts to the Asian side of the Arctic Ocean and the North American trough is replaced by a blocking ridge, high-arctic glaciers experience high melt (Alt, 1987).

7.3.3. Permafrost

The direct influence of climate on permafrost includes the effects of air temperatures and solar radiation. Permafrost also is influenced indirectly by local factors, many of which also have a climate component. These include the thickness and duration of snow cover, the type of vegetation, the properties of the organic layer and soil, and the characteristics of running water (Harris and Corte, 1992; Schmitt, 1993). These factors interact in complex ways, making it difficult to asses the sensitivity of permafrost, and its response, to climate change (Koster, 1991, 1994; Koster and Nieuwenhuijzen, 1992; Koster *et al.,* 1994; Nelson and Anismov, 1993). However, climatic warming usually causes an increase in the thickness of the active layer via melting at the permafrost table (the active layer is the zone of annual freezing and thawing above permafrost). Surface disturbances appear in the first few years, but changes in temperature profiles within the permafrost may be delayed by decades to centuries. The response of permafrost is very dependent on initial ground temperatures and the latent heat (Geo-engineering Ltd, 1995). Displacement of the permafrost base—the final response—takes years to millennia, depending on the depth, thickness, and conductivity of the Earth material. Discontinuous permafrost tends to be most sensitive to climate change because it is usually within one or two degrees of 0°C. Many surface processes that preserve permafrost also are affected by climate change.

Depending on initial conditions, even small changes in climatic regimes cause permafrost to thaw. Simulations performed by Riseborough and Smith (1993) suggest that the rate of thaw and ultimate disappearance of a relatively thin (4.5 m) permafrost profile is highly dependent on the interannual variability of temperature. The broad sensitivity of permafrost to climate change is documented in the former Soviet Union and China, where the permafrost distribution changed substantially during warmer periods of the Quaternary (Kondratjeva *et al.,* 1993; Qiu and Cheng, 1995). The southern limit of lowland permafrost moved at 60 km/°C (Cui and Xie, 1984). In the alpine permafrost of Tibet, the lower elevation changed by 160 m/°C (Cui, 1980). On the northern slope of the Himalayas, the sensitivity is about 80 m/°C (Xie, 1996).

7.3.4. River and Lake Ice

Ginzburg *et al*. (1992) and Soldatova (1993) show that ice formation and break-up correlate with air temperature in the preceding autumn and spring months but not with winter temperatures. Spring warming is more important to the timing of break-up than the overall winter severity and peak ice thickness. Assessing ice chronology and seasonal temperature data from Scandinavia, Canada, and the FSU leads to an estimated sensitivity of both freeze-up and break-up of 5 days/°C. The sensitivity for the length of the ice season is about 10 days/°C.

Freeze-up and ice-loss dates for lakes correlate with autumn and winter air temperatures. Assel and Robertson (1995) show that the sensitivity of freeze-up dates for the Great Lakes is approximately 7 days/°C. For sixty-three smaller lakes in Finland, Palecki and Barry (1986) suggest 5.5 days/°C. Anderson *et al*. (1995) show that interannual variation in ice break-up for twenty U.S. lakes can be explained by ENSO.

7.3.5. Sea Ice

The thickness of fast ice—which grows in fjords, bays, and inlets in the Arctic; along the open coast in shallow water; and in channels of restricted dimensions—is correlated to the number of degree-days of freezing since the beginning of winter. In pack ice, however, the relationship between ice extent or ice thickness and temperature is less clear. Chapman and Walsh (1993) report that there is a statistically significant decrease in Arctic sea-ice extent in winter correlated with atmospheric warming, but there are no trends in winter Arctic ice extent or in the Antarctic.

The sensitivity of oceanic sea-ice cover to climate change is not well-understood. From the data comparisons made so far by McLaren (1989), Wadhams (1989, 1990a), and McLaren *et al*. (1992), the following tentative conclusions can be drawn.

Ice reaching the Fram Strait via the Trans Polar Drift Stream along routes where it is not heavily influenced by a downstream land boundary shows great consistency in its mean thickness from season to season and from year to year, at latitudes from 84°30'N to 80°N and in the vicinity of 0° longitude. Ice upstream of the land boundary of Greenland shows great changes in mean ice draft due to anomalies in the balance between pressure-ridge formation through convergence and open-water formation through divergence. The overall extent of sea ice is largely determined by ice transport via currents and wind and is not necessarily directly related to *in situ* freezing and melting controlled by temperature (Allison, Brandt, and Warren, 1993; Allison and Worby, 1994). As yet, there is no conclusive evidence of systematic thermodynamic thinning of the sea-ice cover, as might be caused by global warming.

Ocean circulation systems—such as the Atlantic conveyor belt—are thought to be sensitive to changes in sea-ice export from the Arctic (Aagaard and Carmack, 1989). Salt flux from local ice production plays an important role in triggering narrow convective plumes in the central gyre region of the Greenland Sea in winter (Rudels, 1990). Here frazil and pancake ice production due to cold-air outbreaks from Greenland can yield high salt fluxes. There already is evidence that a reduction in ice production has produced an ocean response in the form of reduced volume and depth of convection (Schlosser *et al.,* 1991).

7.4. What Will Be the Impact of Future Climate Change on the Cryosphere?

Outputs from three Transient General Circulation Models (GCMs) in Greco *et al*. (1994) were applied to the major cryospheric regions of the world in order to assess impacts for the decade about 2050. They predict that most regions of the cryosphere will warm by 0.5–2.5°C, but some will be wetter and others drier. The amount of warming is consistent among

Box 7-1. Ice Sheets

The great ice sheets of Antarctica and Greenland have changed little in extent during this century. Dynamic response times of ice sheets to climate change are on the order of thousands of years, so they are not necessarily in equilibrium with current climate. Observational evidence for Antarctic and Greenland ice sheets is insufficient to determine whether they are in balance or have decreased or increased in volume over the last 100 years (see Chapter 7, *Changes in Sea Level*, in the Working Group I volume). Zwally *et al*. (1989) note that the Greenland ice sheet surface elevation is increasing by 0.23 m/yr, but this issue of stability remains contentious (Douglas *et al.,* 1990; Jacobs, 1992). There is evidence for recent increases in snow accumulation in East Antarctica (Morgan *et al.,* 1991). On the other hand, Kameda *et al*. (1990) report that the thickness of this ice sheet has decreased by about 350 m during the last 2,000 years.

If Antarctica were to warm in the future, its mass balance would be positive (see Chapter 7, *Changes in Sea Level*, in the Working Group I volume). The rise in temperature would be insufficent to initiate melt but would increase snowfall. Concern has been expressed that the West Antarctic ice sheet may "surge." Working Group I projects that the probability of this occuring within the next century "may be relatively remote, but not zero." There is considerable doubt regarding the possible dynamic response of ice sheets.

The response of Greenland to warming is likely to be different. Both the melt rates at the margins and the accumulation rates in the interior should increase. The former rate is expected to dominate. Thus, the mass balance would become negative as temperatures rise. Several studies have examined precipitation variations over Greenland (e.g., Bromwich *et al*., 1993; Kapsner *et al*., 1995). The likely changes in atmospheric circulation and moisture flux (Calanca and Ohmura, 1994) must be determined before future snowfall can be predicted. Surface topography also plays an important role in determining regional accumulation on the Greenland ice sheet (Ohmura and Reeh, 1991).

Most sheets descend to sea level, where they produce icebergs as pieces break off and float away. These endanger shipping. In Greenland, most iceberg calving comes from a relatively small number of fast-moving tidewater glaciers (Reeh, 1989). Antarctic ice shelves are much larger and grow for periods of 20 to 100 years, regularly calving only relatively small icebergs before enormous pieces break off (Orheim, 1988). For example, giant icebergs have calved from the Filchner, Larsen, Ross, and Shackleton Ice Shelves in recent years. The northerly Wordie Ice Shelf may be systematically breaking up (Doake and Vaughan, 1991), and there have been large changes to the Larsen Ice Shelf culminating in recent large calving rates (Skvarca, 1993, 1994)—including the breakout of a vast iceberg in February 1995. Despite these findings, there is no clear consensus as to whether the frequency of icebergs, and their danger to shipping, will change with global warming.

the three GCMs in most, though not all, cryospheric regions, but predictions for precipitation are discordant among the three GCMs and must be considered less reliable. Assessments of impacts in this chapter generally are made using output from the UKTR model, mainly because of the better resolution of maps in Greco *et al*. (1994) over alpine and polar regions. See Box 7-1 for a discussion of the possible impacts of climate change brought about by the interaction of ice sheets and sea-level change.

7.4.1. Snow

Continental snow cover will be diminished in extent, duration, and depth by the UKTR climate scenarios in Greco *et al*. (1994). Winter snowlines could move further north by 5–10° latitude. The snow season could be shortened by more than a month, depending on snow depth. In North America, climatic warming based on the Boer *et al*. (1992) scenario would cause a 40% decrease in snow-cover duration over the Canadian Prairies and a 70% decrease over the Great Plains (Brown *et al.,* 1994). Using CCM1 model output for the whole Northern Hemisphere, the area of seasonal snow in February may diminish by 6–20%, with a mean decrease of 12 x 10^6 km^2 (Henderson-Sellers and Hansen, 1995).

Snowfall will begin later and snowmelt will be earlier than at present, so the snow-free season will be extended. The advance in melt time is likely to be less pronounced in the High Arctic, where the snow is so cold that it requires significant warming to produce consistent melt (more than 10°C at many locations). More frequent periods of open water for rivers, lakes, and seas will produce greater snowfall downwind. This will be important near Hudson Bay, the Great Lakes, the Barents Sea, and the Sea of Okhotsk. In the Antarctic, summer temperatures are so low that the present regime of little or no snowmelt will persist.

In alpine areas, the snow line could rise by 100–400 m, depending on precipitation. Higuchi (1991) shows that as warming occurs for many Asian mountains, there is a tendency for rainfall to occur at the expense of snowfall, although the extent of this shift depends on location. Less snow will accumulate at low elevations, but there may be more above the freezing level from any increased precipitation. Martinec *et al*. (1994) and Rango and Martinec (1994) have examined the behavior of a snowmelt-runoff model in various catchments for different climate scenarios. With a rise of 1°C, snow cover would be depleted in winter due to conversion of precipitation to rainfall and increased snowmelt. Five days into the melt season, snow depth would be depleted to the equivalent of 9 days under the present climate. It is estimated that for the New Zealand Alps, total water stored as snow would be reduced by 20% of present for a climate scenario that is 2°C warmer but 10% wetter than present (Garr and Fitzharris, 1994). Impacts of best- and worst-case scenarios on snow cover in the Victorian Alps (Australia) are presented in CSIRO (1994). Simulated average snow cover and the frequency of years with more than 60 days of cover decline at all sites. For the worst case, snow-cover duration at even the highest sites (1,900 m) is halved by 2030 and is near zero by 2070.

7.4.2. Ice Caps and Glaciers

Empirical and energy-balance models both indicate that a large fraction (about one-third to one-half) of presently existing mountain glacier mass could disappear with anticipated warming over the next 100 years (Kuhn, 1993; Oerlemans and Fortuin, 1992). By 2050, up to a quarter of mountain glacier mass could have melted. The scenarios of Greco *et al*. (1994) indicate that some mountain areas will experience an increase in precipitation. Because models demonstrate that increases in temperature usually dominate changes in precipitation, mass balances of glaciers will become negative rather than positive. Glaciers are likely to shrink even where mountains become wetter. An upward shift of the equilibrium line by some 200 to 300 m and annual ice thickness losses of 1 to 2 m are expected for temperate glaciers.

Many mountain chains will lose major parts of their glacier cover within decades. Haeberli and Hoelzle (1995), who have developed algorithms for analyzing glacier inventory data, show that glacier mass in the European Alps could be reduced to a few percent within decades if current warming continues. Nevertheless, the largest alpine glaciers—such as those found around the Gulf of Alaska and in Patagonia, Karakoram, Pamir, Tien Shan, and the Himalayas—should continue to exist into the 22nd century. At high altitudes and high latitudes, glaciers and ice caps may change little in size, but warming of cold firn areas will be pronounced. Their mass balance may be affected through enhanced ablation at low altitudes, while accumulation at higher zones could increase. As an analog, Miller and de Vernal (1992) report that some Arctic glaciers did not shrink during warmer parts of the Holocene but actually grew due to increased precipitation.

There will be pronounced alterations to glacier melt runoff as the climate changes. Glaciers will provide extra runoff as the ice disappears. In most mountain regions, this will happen for a few decades, then cease. For those with very large glaciers, the extra runoff may persist for a century or more and may provide a substantial increase to regional water resources. As the area of ice eventually diminishes, glacial runoff will wane. Tentative estimates have been made for Central Asia (Kotlyakov *et al.,* 1991) based on mass balances from a small number of Tien Shan glaciers for the period 1959–1992. Extrapolation to the whole of Central Asia suggests glacier mass has decreased by 804 km^3 over that time, representing a 15% increase in glacial runoff. Projections to 2100 are presented in Table 7-2; these data assume that glacial runoff varies as a linear function of glacier area.

An alternative approach is based on calculations of ablation and equilibrium-line altitude, according to given climatic scenarios (Kotlyakov *et al.,* 1991; Glazyrin *et al.,* 1990; Dikikh and Dikikh, 1990). Ablation intensifies in Central Asia with

Table 7-2: *Present and possible future extent of glaciation and glacial runoff in Central Asia (after Kotlyakov et al., 1991).*

Component	Present	2100
Area of glaciers (km^2)	115,000	80,000
Decrease of glacier mass (km^3/yr)	25	18
Runoff due to precipitation (km^3/yr)	73	50
Glacial runoff (km^3/yr)	98	68

climate warming because conditions become even more continental. Annual mass balance decreases. These estimates give a high rate of glacier degradation and large alteration of runoff (Figure 7-5). By 2050, the volume of runoff from glaciers in Central Asia is projected to increase threefold.

7.4.3. Permafrost

Anisimov and Nelson (1995) have compiled global permafrost maps for the present day and for a 2050 scenario (see Figure 7-3b). The compilation uses a predictive climate-based permafrost model, in which permafrost is classified on the basis of "surface frost index" (Nelson and Anisimov, 1993) and which also considers the influence of snow cover on the soil thermal regime. Calculated contemporary boundaries, based on the Global Ecosystem Database (GED), show good agreement with Figure 7-3a. Table 7-3 provides the areal extent of permafrost distribution in the Northern Hemisphere for two climate scenarios (GED contemporary and 2050). A 16% shrinkage in total permafrost area is projected by 2050. Subsurface conditions may not be in equilibrium with the surface area shown. Some may take hundreds of years to respond. The extent of permafrost zones in the Southern Hemisphere, which cover only approximately 0.5 x 10 ^{6} km^2 in total, also is likely to decrease.

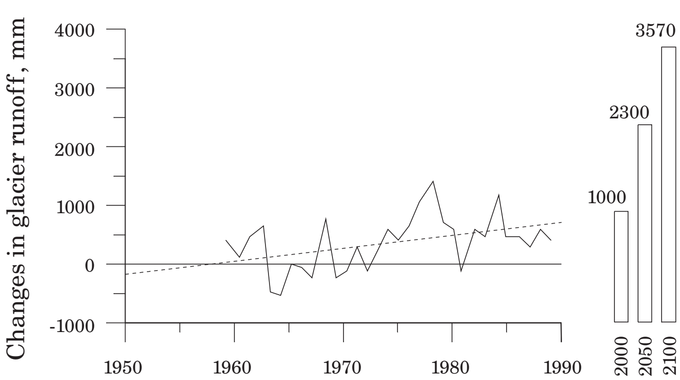

Figure 7-5: Changes in recent and future runoff from Tuyuksu glacier, Zailiyski Alatau, Central Asia (after Kotlyakov *et al*., 1991). Values represent runoff due to ice loss, expressed as departures from long-term mean.

There will be poleward shift of discontinuous and continuous permafrost zones (Woo *et al.,* 1992; Nelson and Anisimov, 1993). These estimates only treat degradation of near-surface permafrost; more attention must be given to the role of latent heat associated with ablation of ground ice. Simulations conducted by Riseborough (1990) indicate that areas with abundant ground ice, such as western Siberia (Burns *et al.,* 1993), may retain substantial amounts of permafrost that are not in equilibrium with new temperature conditions imposed at the surface. The relict permafrost of the southern part of West Siberia, which may well date from the last glacial epoch, is a present-day example of resistance to changes of temperature in deep-seated, ice-rich permafrost (Kondratjeva *et al.,* 1993).

In areas where permafrost is discontinuous, long-term warming ultimately will lead to its thinning and general disappearance (Wright *et al.,* 1994). Long-term temperature measurements in discontinuous permafrost in Alaska indicate that an increase in surface temperature of 2°C would cause most of the permafrost south of the Yukon River and on the south side of the Seward Peninsula to thaw. In continuous permafrost, a warming of the permafrost, thickening of the active layer, and changes in thaw lake dynamics are likely. Initial degradation will be contemporaneous with the alteration of the climatic signal (Osterkamp, 1994; Kane *et al.,* 1991), but as the thaw plane penetrates more deeply, lag times will increase.

Xie (in press) estimates that the southern boundary of permafrost in northeast China could be at 48°N latitude by the year 2100, and the predominantly continuous permafrost zone may recede to 52°N—similar to that in the Climatic Optimum of the Holocene (Lu Guowei *et al.,* 1993). In Tibet, it is estimated that an air-temperature warming of 3°C would raise the permafrost limit to an elevation of 4,600 m. Continuous frozen ground would disappear, except in the northwestern part of the plateau and areas around the Fenghuo Shan, in the event of a 5°C rise in temperature (Xie, in press).

In the FSU and Europe, reconstructions of the permafrost zone in past warm and cold periods are used as palaeoanalogues (Velichko and Nechaev, 1992; Vanderberghe, 1993). The most significant change in the thermal regime will occur in the high

Table 7-3: *Calculated contemporary and future areas of permafrost in the Northern Hemisphere (10^6 km^2) (based on Nelson and Anisimov, 1993).*

Zone	Contemporary	2050	% Change
Continuous Permafrost	11.7	8.5	-27
Discontinuous Permafrost	5.6	5.0	-11
Sporadic Permafrost	8.1	7.9	-2
Total	**25.4**	**21.4**	**-16**

latitudes within the present-day zones of the tundra and boreal forest. With a rise in temperature of 1°C, the permafrost would be partially preserved to the east of the Pechora river. In the south of Yamal and Gydan, discontinuous permafrost would prevail. Continuous permafrost would be restricted to north of 70°N. It would remain to the east of the Yenisey at the same latitudes. The active layer of fine-grained soils could increase by 20–30 cm. With a temperature rise of 2°C, continuous permafrost would disappear in the north of Europe. From the Lower Ob to the Lower Hatanga, as well as in the Anadyr lowland, only island and discontinuous permafrost would survive. Continuous permafrost would exist only in the lowlands of the Taimyr Peninsula, Lena, lower Kolyma, and lower Indigirka basins. The stratum of seasonal thawing would increase by 40–50 cm. These changes will result in the activation of solifluction, thermokarst and thermoerosion processes, an increase in bogs, and alterations to large tracts of vegetation—which in turn will alter food sources for traditional tundra species within permafrost areas.

Other mathematical models assess permafrost dynamics during warming (MacInnes *et al.,* 1990; Burgess and Riseborough, 1990; Romanovsky, Maximova, and Seregina, 1991; Romanovsky, Garagula, and Seregina, 1991; Osterkamp and Gosink, 1991; Nixon and Taylor, 1994; Nixon *et al.,* 1995). Changes in permafrost temperatures and thaw depths under an increase in annual average air temperatures of 2 and 4°C are simulated by Vyalov *et al*. (1993). Results indicate that warming in the tundra will produce a slight increase in thawing depths, with no radical changes in permafrost conditions. Further south, more marked effects could occur—with a shift to discontinuous permafrost, formation of taliks, and degradation of ground ice. If warming is long-lasting, the permafrost line would recede northward by 500 km along 70°E longitude and by 1,200 km along 100°E longitude. Given an average thawing rate of 100 mm/yr, thawing of the upper 10 m would take a century. Lower limits of permafrost in mountain areas could rise by several hundred meters, although this is dependent on future snow depths (Hoelzle and Haeberli, 1995). Owing to the slow reaction of thermal conditions at depth, pronounced disequilibria are most likely to result over extended time periods and wide areas.

Ice-saturated permafrost forms an impervious layer to deep infiltration of water, maintaining high water tables and poorly aerated soils. Significant increases in active layer depth or loss of permafrost is expected to cause drying of upper soil layers in most regions, as well as enhanced decomposition of soil organic matter. Loss of a sizable portion of more than 50 Gt of carbon in Arctic soils and 450 Gt of carbon in soils of all tundra ecosystems could cause an appreciable positive feedback on the atmospheric rise of carbon dioxide. Marion and Oechel (1993) have examined Holocene rates of soil carbon change along a latitudinal transect across arctic Alaska and conclude that arctic areas will continue to act as a small sink for carbon.

Recent warming and drying apparently have shifted arctic ecosystems from carbon sequestration (as occurred during the Holocene and historical past) to carbon dioxide loss to the atmosphere (Oechel *et al.,* 1993). Estimates of present losses of carbon dioxide from arctic terrestrial ecosystems to the atmosphere range from 0.2 Gt of carbon (Oechel *et al.,* 1993) to much higher values in winter for the Russian Arctic (Zimov *et al.,* 1993; Kolchugina and Vinson, 1993). Malmer (1992) estimates that about 25% or less of the carbon may be released in the form of methane (CH_4). A climatic shift to warmer temperatures in the future would increase the release of CH_4 from deep peat deposits, particularly from tundra soils. It is expected that the release of CO_2 would increase, though not by more than 25% of its present level (Malmer, 1992). Wetter soils could lead to increased methane loss, and a drier tundra might become a sink for atmospheric methane (Christensen, 1991, 1993; Fukuda, 1994).

Large amounts of natural gas—mostly methane—are stored in the form of gas hydrates, although their distribution is not well-known. In the Canadian Arctic Islands and Beaufort-Mackenzie region, analysis of thermal and geophysical logs indicates that 2–4 x 10^3 Gt of methane is stored as hydrates. Decomposition is occurring presently beneath the shelves of the Arctic Ocean in response to the increase of surface temperatures accompanying the recent marine transgression. For the Beaufort Shelf, an estimated 10^5 m^3/km^2 may decompose over the next century (Judge and Majorowicz, 1992; Judge *et al.,* 1994).

There are very few data for the methane content of permafrost itself. Samples obtained near Fairbanks (Kvenvolden and Lorenson, 1993) and the Prudhoe Bay area (Moraes and Khalil, 1993; Rasmussen *et al.,* 1993) suggest substantial variability. Dallimore and Collett (1995) found high methane concentrations in ice-bonded sediments and gas releases suggest that pore-space hydrate may be found at depths as shallow as 119 m. This raises the possibility that gas hydrates could occur at much shallower depth and be more rapidly influenced by climate change than previously thought. Fukuda (1994) estimates methane emissions from melting ground ice in northern Siberia cover the range 2–10 Mt/yr.

7.4.4. River and Lake Ice

Under conditions of overall annual warming, the duration of river-ice cover would be reduced through a delay in the timing of freeze-up and an advancement of break-up (Gray and Prowse, 1993). For freeze-up, higher water and air temperatures in the autumn would combine to delay the time of first ice formation and eventual freeze-up. If there also is a reduction in the rate of autumn cooling, the interval between these two events also will increase. The frequency and magnitude of major frazil-ice growth periods could be reduced. This may alter the types of ice that constitute the freeze-up cover and has implications for hydrotechnical problems associated with particular ice forms.

Many rivers within temperate regions would tend to become ice-free or develop only intermittent or partial ice coverage. Ice growth and thickness would be reduced. In colder regions, the present ice season could be shortened by up to a month by 2050. Warmer winters would cause more mid-winter break-ups as rapid snowmelt, initiated particularly by rain-on-snow events, becomes more common. Warmer spring air temperatures may affect break-up severity, but the results would be highly site-specific because break-up is the result of a complex balancing between downstream resistance (ice strength and thickness) and upstream forces (flood wave). Although thinner ice produced by a warmer winter would tend to promote a thermal break-up, this might be counteracted to some degree by the earlier timing of the event, reducing break-up severity (Prowse *et al.,* 1990).

Changes in the size of the spring flood wave depend on two climate-related factors: the rate of spring warming and the water equivalent of the accumulated winter snowpack. Whereas greater and more-rapid snowmelt runoff favors an increase in break-up severity, the reverse is true for smaller snowpacks and more protracted melt. The final effect on break-up also will depend on the potentially conflicting roles of ice strength and thickness.

For arctic lakes, the duration of ice cover would be shortened (Assel *et al.,* 1995). A longer open-water period, together with warmer summer conditions, would increase evaporative loss from lakes. Some patchy wetlands and shallow lakes owe their existence to a positive water balance and the presence of an impermeable permafrost substrate that inhibits deep percolation. Enhanced evaporation and ground thaw would cause some to disappear. Using a 20-year period of record, Schindler *et al.* (1990) found that climate warming increased the length of the ice-free season by 3 weeks, as well as having numerous indirect effects (see Chapter 10).

If the warming of the 1980s continues unabated over the next 10–20 years, ice cover on the Great Lakes of North America will likely be similar to or less than that during the 1983 ice season, which was one of the mildest winters of the past 200 years. Mid-lake ice cover did not form, and ice duration and thickness were less than normal (Assel *et al.,* 1985). Complete freezing will become increasingly infrequent for larger and deeper embayments in the Great Lakes. The duration of ice cover will decrease as freeze-up dates occur later and ice-loss dates occur earlier. Winters without freeze-up will begin to occur at small inland lakes in the region. Winter lake evaporation also may increase due to the decreased ice cover.

7.4.5. Sea Ice

GCM experiments with simplified treatments of sea-ice processes predict large reductions in sea-ice extent but produce widely varying results and do not portray extent and seasonal changes of sea ice for the current climate very well. Boer *et al.* (1992) estimate that with a doubling of greenhouse gases, sea ice would cover only about 50% of its present area. CCM1 model output presented in Henderson-Sellers and Hansen (1995) projects a 43% reduction for the Southern Hemisphere and a 33% reduction for the Northern Hemisphere. The global area of sea ice is projected to shrink by up to 17 x 10^6 km^2.

Using empirical ice growth-melt models, Wadhams (1990b) predicts that in the Northwest Passage and Northern Sea Route, a century of warming would lead to a decline in winter fast-ice thickness from 1.8–2.5 m at present to 1.4–1.8 m and an increase in the ice-free season of 41–100 days. This effect will be of great importance for the extension of the navigation season in the Russian Northern Sea Route and the Northwest Passage. A possible feedback with snow thickness may alter these relationships. As Arctic warming increases open-water area, precipitation may increase and cause thicker snow cover. The growth rate of land fast ice will decrease, as has been directly observed (Brown and Cote, 1992). But if snow thickness is increased to the point where not all ice is melted in summer, then the protection that it offers the ice surface from summer melt could lead to an increase, rather than decrease, in equilibrium ice thickness.

Predicting future thickness of moving pack ice is a difficult problem, because dynamics (ocean and wind currents), rather than thermodynamics (radiation and heat components), determine its area-averaged mean thickness. Wind stress acting on the ice surface causes the ice cover to open up to form leads. Later under a convergent stress, refrozen leads and thinner ice elements are crushed to form pressure ridges. Exchanges of heat, salt and momentum are all different from those that would occur in a fast ice cover. The effects of variable thickness are very important. The area-averaged growth rate of ice is dominated (especially in autumn and early winter when much lead and ridge creation take place) by the small fraction of the sea surface occupied by ice less than 1 m thick. In fast ice, climatic warming will increase sea-air heat transfer by reducing ice growth rates. However, over open leads a warming will decrease the sea-air heat transfer, so the area-averaged change in this quantity over moving ice (hence its feedback effect on climatic change itself) depends on the change in the rate of creation of new lead area. This is itself a function of a change in the ice dynamics, either driving forces (wind field) or response (ice rheology). Hibler (1989) pointed out a further factor relevant to coastal zones of the Arctic Ocean, such as off the Canadian Arctic Archipelago, where there is net convergence and the mean thickness is very high (7 m or more) due to ridging. Here the mean thickness is determined by mechanical factors, largely the strength of the ice, and is likely to be insensitive to global warming.

Ice dynamics also have other effects. In the Eurasian Basin of the Arctic the average surface ice drift pattern is a current (the Trans Polar Drift Stream) which transports ice across the Basin, out through Fram Strait, and south via the East Greenland Current into the Greenland and Iceland Seas, where it melts. The net result is a heat transfer from the upper ocean in sub-Arctic seas into the atmosphere above the Arctic Basin. A change in area-averaged freezing rate in the Basin would thus cause a change of similar sign to the magnitude of this long-range heat transport.

An identical argument applies to salt flux, which is positive into the upper ocean in ice growth areas and negative in melt areas. Salt is transported northward via the southward (Lemke *et al.,* 1990) ice drift. A relative increase in area-averaged melt would cause increased stabilization of the upper layer of polar surface water, and hence a reduction in heat flux by mixing across the pycnocline. A relative increase in freezing would cause destabilization and possible overturning and convection.

If ice were to retreat entirely from the central gyre region of the Greenland Sea, it may cause deep convection to cease. Already there is evidence from tracer studies (Schlosser *et al.,* 1991) of a marked reduction in the renewal of the deep waters of the Greenland Sea by convection during the last decade. It is not known whether this is part of a natural variation, or a response due to greenhouse warming. If continued, it could have a positive feedback effect on global warming, since the ability of the world ocean to sequestrate carbon dioxide through convection is reduced. Given the complexity of these interactions and feedbacks, it is not at all clear what the overall effect of an air temperature increase on the Arctic ice cover and upper ocean will be. Sensitivity studies using coupled ocean-ice-atmosphere models are required, but results are not yet available.

In the Antarctic, where the sea ice cover is divergent and where land boundaries are less important, it is more reasonable to suppose that the main effect of global warming will be a simple retreat of the ice edge southward. However, Martinson (1990) has demonstrated that even here, a complex set of feedback mechanisms comes into play when the air temperature changes. The balance of lead concentration, upper ocean structure and pycnocline depth adjusts itself to minimize the impact of changes, tending to preserve an ice cover even though it may be thinner and more diffuse.

7.5. What Will Be the Impact of These Cryospheric Changes?

These changes in the cryosphere will affect many natural and managed ecosystems, as well as socioeconomic systems. Only the most notable are considered here. Two regions will be most affected: Temperate mountains and tundra lands (see Boxes 7-2 and 7-3, respectively).

7.5.1. Impact on Hydrology and Water Resources

The impact of climate change on water resources in alpine areas can be large, as shown by Gleick (1987a, 1987b), Martinec and Rango (1989) and Moore (1992), and are discussed at length in Chapter 10. Seasonal changes can be

Box 7-2. Impact on Temperate Mountains

Revegetation of terrain following deglaciation is slow in high-mountain areas. This leaves morainic deposits unprotected against erosion for extended time periods (decades to centuries). There will be increased sediment loads in alpine rivers and accelerated sedimentation in lakes and artificial reservoirs at high altitude. On slopes steeper than about 25–30 degrees, stability problems, such as debris flows, will develop in freshly exposed or thawing non-consolidated sediments.

At places of pronounced glacier retreat, changes in stress distribution and surface temperature conditions in rock walls of deeply cut glacier troughs must also be anticipated, so that massive rock slides will occur in the deglaciated valley (Clague and Evans, 1992, 1994; Evans and Clague, 1993, 1994). Steep hanging glaciers which are partially or entirely frozen to their beds could become less stable. On the other hand, some steep glacier tongues with present-day potential for large ice avalanches will disappear. Lakes dammed by landslides, moraines and glaciers can drain suddenly and produce floods or debris flows orders of magnitude larger than normal stream flow. Processes related to ice retreat such as glacier avalanches, slope instability caused by debuttressing, and glacier floods from moraine or ice-dammed lakes may pose hazards to people, transport routes, and economic infrastrucure in mountain areas. The general tendency in high mountains will be an upslope shifting of hazard zones and widespread reduction in stability of formerly glaciated or perennially frozen slopes (Barsch, 1993; Haeberli, 1992; Gu *et al.,* 1993; Dutto and Mortara, 1992).

Shrinkage of permafrost and snow cover will eliminate snow metamorphism, and avalanche formation in high mountain regions and adjacent lowlands (Keller and Gubler, 1993; Tenthorey, 1992). An increased frequency of catastrophic snow avalanches may occur if sudden cold spells and a mixture of cold and warm air masses are more common in late winter, or if periods of rapid warming are accompanied by heavy rain follow periods of intensive snowfalls. There is insufficient evidence to indicate whether the frequency of such events will change. Avalanches are expected to be less of a hazard than at present.

The empirical basis for assessing alpine hazard probabilities comes from historical documents, statistics of measured time series or traces in nature of past events with long recurrence intervals. A future problem is that these will lose more and more of their significance as climate changes. This is because floods, avalanches, debris flows, and rock falls could have different magnitudes and frequencies than in the present climate, as the condition of the cryosphere in high mountain areas evolves beyond the range of Holocene and historical variability.

Box 7-3. Impact on Tundra Lands

A critical factor influencing the response of tundra to warming depends on the presence of ground ice. Ground ice is generally concentrated in the upper 10 meters of permafrost, the very layers that will thaw first as permafrost degrades. This loss is effectively irreversible, because once the ground ice melts, it cannot be replaced for millenia, even if the climate were to subsequently cool. Response of the permafrost landscape to warming will be profound, but will vary greatly at the local scale, depending on detail of ground ice content. As substantial ice in permafrost is melted, there will be land subsidence. This process of thermokarst erosion will create many ponds and lakes and lead to coastal retreat.

A forerunner of future landscapes can be seen in areas of massive ground ice, such as in Russia, where past climatic warming has altered the landscape by producing extensive flat-bottomed valleys. Ponds within an area of thermokarst topography eventually grow into thaw lakes. These continue to enlarge for decades to centuries, due to wave action and continued thermal erosion of the banks. Liquefaction of the thawed layer will result in mudflows on slopes in terrain that is poorly drained or that contains ice-rich permafrost. On steeper slopes there will also be landslides (Lewkowicz, 1992). Winter discharge of groundwater often leads to ice formation, and this is expected to increase on hillslopes and in the stream channels of the tundra.

Changes in landscape, sea-ice distribution, and river and lake ice could have a major impact on indigenous people who live in Arctic regions and depend upon traditional occupations, food gathering, and hunting (Kassi, 1993; Roots, 1993; Wall, 1993). These include the Inuit of North America and Greenland and the various reindeer herding groups of Eurasia. They depend directly on the living resources of the area and often travel on ice, so their livelihood may be widely affected. Ice roads and crossings are commonly used to link northern settlements. The greatest economic impact is likely to stem from decreases in ice thickness and bearing capacity, which could severely restrict the size and load limit of vehicular traffic (Lonergan *et al.,* 1993). There is likely to be a change in the migration patterns of polar bears and caribou, along with other biological impacts.

There will be considerable impacts of climate change on resource management in the tundra (Wall, 1993). Some infrastructure and mining activities and structures will be threatened by thawing of permafrost. Water resources will change in that the seasonality of river flows will be different. Environmental changes are expected to be greater than for many other places on Earth (Roots, 1993).

marked when the cryosphere is involved in river flow. In their snowmelt runoff simulations, Martinec *et al*. (1994) report that for a 4°C rise in temperature of the Rio Grande basin in the USA, winter runoff increases from 14% of the total annual flow at present to 30%. Summer runoff decreases from 86% of annual flow to 70%. Such changes in snow runoff can affect irrigation water and electricity supply.

Less snow and glacier ice will influence the seasonality of river flow by reducing meltwater production in the warm season. The expected smoothing of the annual runoff amplitude could be both beneficial (e.g., energy production in winter, reduction of summer flood peaks) and adverse (e.g., reduced water supply for summer irrigation in dry areas, more frequent winter floods). As mountain glaciers begin to disappear, then eventually the volume of summer runoff will be reduced due to the loss of ice resources. Consequences for downstream agriculture, which relies on this water for irrigation, are in some places very unfavourable. For example, low and midland parts of Central Asia are likely to gradually change into a more arid, interior desert.

As climatic warming occurs, there will be notable changes in the hydrology of Arctic areas (Woo, 1990). The nival regime runoff pattern will weaken for many rivers in the permafrost region, and the pluvial influence upon runoff will intensify for rivers along the southern margin of the Arctic, regions of Eurasia, and North America. Should climatic change continue, the vegetation will likely be different from today. When the lichens and mosses, that tend to be suppressors of evapotranspiration, are replaced by transpiring plants, evaporative losses will increase. Enhanced evaporation will lower the water table, followed by changes in the peat characteristics as the extensive wetland surfaces become drier (Woo, 1992).

In permafrost regions, increased thawing deepens the active layer, allowing greater infiltration and water storage, especially for rain that falls during the thawed period. Warming of the ground will also lead to the formation of unfrozen zones within the permafrost which provide conduits for groundwater flow. The chemical composition and the amount of groundwater discharge may be changed as sub-permafrost or intra-permafrost water is connected to the surface. In autumn and winter, more groundwater should be available to maintain baseflow, further extending the streamflow season.

Structures such as pipelines, airstrips, community water supply and sewage systems, and building foundations are susceptible to performance problems if existing, frozen foundations or subgrades thaw, even minimally. Special measures would be needed to ensure the structural stability and durability of installations for tourism, mining industry, and telecommunication in permafrost areas affected by climate warming (Anyia *et al.,* 1992; Haeberli, 1992; Vyalov *et al.,* 1993). Transport links could also be affected. For example, the permafrost zone in China contains more than 3,000 km of railway and over 13,000 km of highway. Thawing induced by climate warming will result in serious disruption and increased maintenance costs from ground subsidence, side-slope slumpings, landslides, icings and ice mound growth (Yang, 1985). On the other hand, many northern cities will spend less money on snow and ice clearance. Engineering design criteria will need to be modified to reflect changing snow and frost climates, deepening of the active layer over permafrost, and warming and ultimate disappearance of marginal or discontinuous permafrost. Present permafrost engineering commonly designs for the warmest year in the past 20 years of record (Esch, 1993). Such criteria may need to be reviewed and revised.

Where climate change alters the river-ice regime, substantial effects on the hydrology can be expected that will affect flow, water levels, and storage. For cold continental rivers, many hydrologic extremes, such as low flows and floods, are frequently more a function of ice effects than landscape runoff. At freeze-up, the hydraulic resistance of an accumulating cover can induce sufficiently large hydraulic storage that the river flow falls below that normally expected (Gerard, 1990; Gray and Prowse, 1993). Projected climates will delay the timing of freeze-up and so prolong the autumn low flow period. At break-up, the rapid hydraulic storage and release of water by river-ice jams often forms the most significant hydrologic event of the year. Break-up flood levels typically exceed those that develop during the open-water period, even though actual discharge of water is lower. The impact of climatic warming will be to advance the timing of break, but its effect on break-up flooding is not clear.

Of all river-ice processes, ice-jams are the major source of economic damages, averaging approximately $CDN20–30 million per year in Canada alone (Van Der Vinne *et al.,* 1991). Changes in damages from such events depend on how climate change affects the frequency and severity of river-ice freeze-up and break-up events. Less river ice and a shorter ice season in northward flowing rivers of Canada, Russia and Siberia should enhance north-south river transport. When combined with less sea ice in the Arctic, new opportunites for reorganization of transport networks and trade links will arise. Ultimately those changes could affect trading patterns among Russia, USA, Canada, northern Europe, and Japan.

7.5.2. Impact on the Hydroelectric Industry

Altered future climates could have a significant impact on the seasonal distribution of snow storage, runoff into hydroelectric catchments, and aggregated electricity consumption. Garr and Fitzharris (1994) show that in New Zealand the winter (Austral) gap between electricity consumption and generation is reduced for a climate scenario that might be expected about 2050 (Figure 7-6). The electricity system is made less vulnerable to climate variability. Water supply is increased, but demand, which is largely driven by domestic heating, is reduced. There would be less need for new hydro plants and water storage. On the whole, these amount to net benefits. In countries where heating is supplied by natural gas, these changes will be less obvious. Instead the demand may well increase in the summer due to greater use of air conditioners due to a higher frequency of very hot days.

River ice creates a host of hydrotechnical difficulties for the operation of hydropower facilities, ranging from the blockage of trash racks by frazil ice to the curtailment of operations so as to avoid freeze-up flooding. One Canadian hydropower company estimates revenue losses of over $CDN1 million/year associated with the release of sufficient flow to avoid the downstream freeze-off of a major tributary (Foulds, 1988). Major economic savings are likely if the length of the ice season is reduced by climatic warming. Operational problems during the ice covered period will remain, and could even increase because some rivers may experience a higher frequency of winter break-ups.

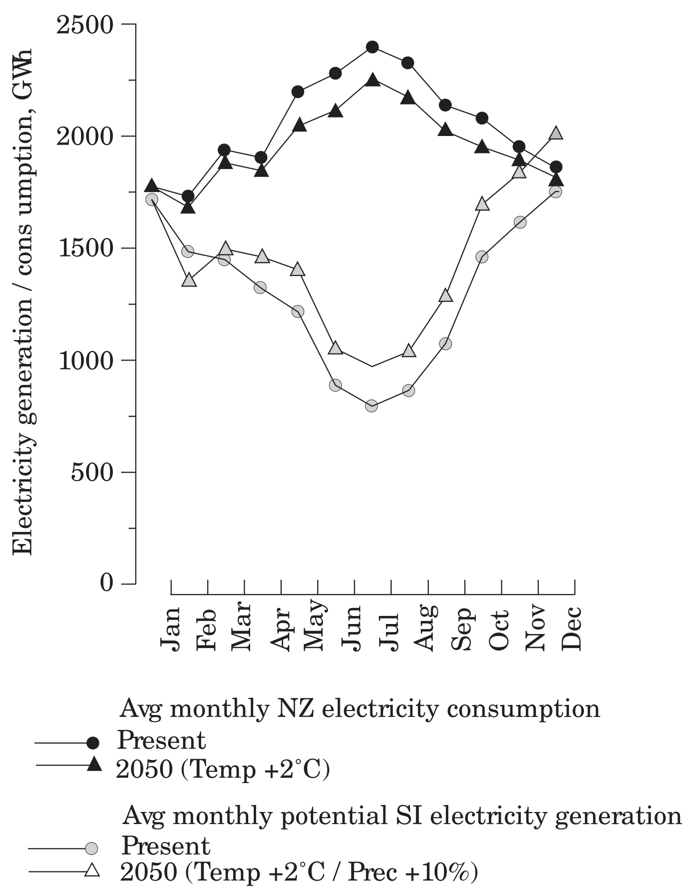

Figure 7-6: Synthesis of hydroelectricity generation and total consumption of electricity for New Zealand: The present compared with 2050 (after Garr and Fitzharris, 1994).

7.5.3. Impact on Shipping

In Canada, absence of sea ice south of Labrador would eliminate Canadian Coast Guard ice breaking requirements. This means an annual saving of between $CDN15–20 million. Even larger savings can also be expected in the FSU. The substantial reductions in sea ice in the Arctic Ocean that are expected to occur with climatic warming will increase opportunities for shipping there and open up new trade routes. The effect of annual warming on ice calving, simulated using a simple degree day model (Brown, 1993) shows that for every 1°C of warming there would be a 1° latitude retreat of the iceberg occurrence in the Atlantic Ocean. In the Southern Ocean, any effects of reduced sea ice will be economically less pronounced.

Ice-breakers keep river channels open for ship traffic within Russia on the major northern rivers, and to a limited degree in North America within the Great Lakes system. Considering the high operational costs of ice-breaking, any reduction in the ice season should translate into significant cost-savings. With warming there will be a longer shipping season, allowing increased passage of goods and services and longer time and larger area open for commercial fishing (Reycraft and Skinner, 1993). Changes in fresh colder water in the North Atlantic may change the distribution of fisheries.

7.5.4. Impact on the Oil Industry

Projected climatic warming could extend summer open water in the Beaufort Sea an additional 200–800 km offshore (McGillivray *et al.,* 1993). The fetch, and frequency of extreme wave heights, will therefore increase. It is estimated that the frequency of 6m waves would rise from 16% to 39% of the time (McGillivray *et al.,* 1993). Present design requirements for long-lived coastal and offshore structures, such as oil installations, will be inadequate under these conditions.

A possible beneficial effect would be shorter winters disrupting construction, exploration, and drilling programs. A decrease in thickness of first-year ice of 50–70% is projected for the Arctic, which will extend the drilling seasons for floating vessels considerably. Costs of drilling "downtime" on offshore oil and gas drilling explorations could practically be eliminated due to iceberg and sea-ice absence, saving more than $CDN40 million annually (Stokoe, 1988).

7.5.5. Impact on the Tourist Industry

Many mountainous regions of the world are used for tourism. A prerequisite for such commercially important activities as skiing is an extended snow cover of sufficient depth. Despite increases in precipitation, warming will decrease the cover, depth, and quality of snow and impair winter tourism in most alpine countries. Subsequent socioeconomic consequences could be detrimental for many mountain communities (Foehn, 1990). In the United States, Cline (1992) estimates ski industry losses from projected warming at $1.7 billion annually. In some countries, ski resorts could be re-established at higher elevations, but this too has associated transportation and emission problems. In others, such as Australia, the ski industry could, under the worst case scenario, be eliminated. Less snow and fewer glaciers on mountains will also diminish the quality of many alpine vistas, where their scenic appeal depends upon their presence in the general landscape. Some countries are still sufficiently high that their mountains will retain snow, so will possess an increasingly scarce, but valuable, scenic and recreational resource.

Reduced sea ice will provide safer approaches for tourist ships and new opportunities for sightseeing around Antarctica and the Arctic.

7.6. What Do We Still Need to Know?

There are many uncertainties in understanding what is currently happening to critical components of the cryosphere, especially ice sheets, sea ice, and permafrost, mainly because existing monitoring systems are inadequate. Critical questions remain about how each component of the cryosphere will react to climate change and a multitude of proposals could be made. The most important are synthesized as follows:

- Climatic scenarios produced by GCMs need to be refined so as to provide more detail for mountain and polar regions. Their output for regions of the globe where snow, ice, and permafrost are dominant needs to be better verified for the present climate. Imprecise estimates of future polar and alpine precipitation, and particularly snow depth, are a major constraint in predicting the behavior of most components of the cryosphere and their impact.
- Research should focus on processes that are driven by interactions between the atmosphere and the cryosphere. Improving understanding of these processes would allow the construction of more sophisticated and realistic climate sensitivity models.
- Monitoring of key components of the cryosphere must continue. The mass balance of the ice sheets of the world is poorly known. Databases need to be further developed and maintained. They provide the benchmark for assessing future change and for model testing. More statistical work is required on existing databases to improve knowledge of cryospheric trends.
- Effects of cryospheric change on other natural systems need to be better understood and quantified, particularly where they affect human communities and economic systems of agriculture, forestry, tourism, transport, and engineered structures. Future response of Arctic sea ice is especially critical because large changes will have profound climatic, economic, trade, and strategic implications.
- New methods are needed to assess probabilities of natural hazard and risk that take into account changing

climate. For hazard mitigation measures in mountain environments of developing countries, transfer of technologies for preparing the necessary assessments would be helpful.

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