CHAPTER 6

Cryospheric Systems

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Chapter 6 Contents

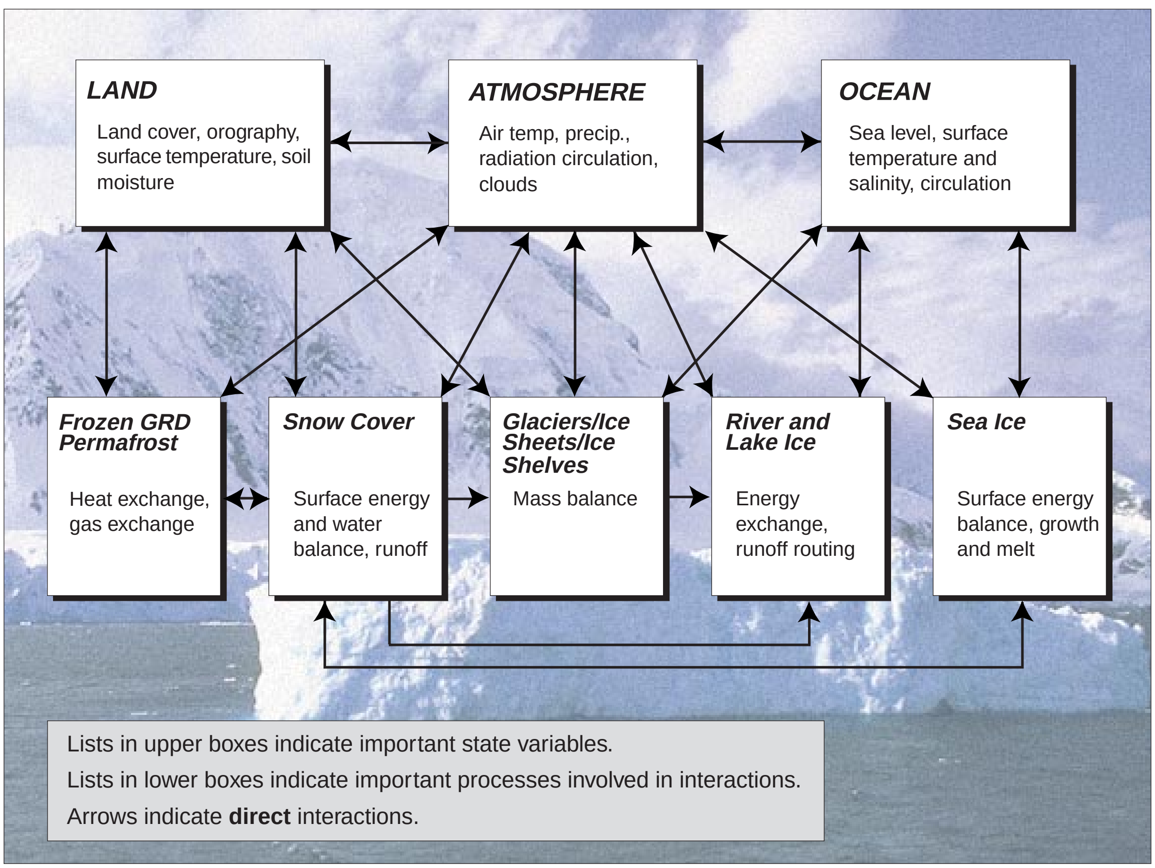
6.1 Overview

6.1.1 Rationale for studying the cryosphere

The term "cryosphere" traces its origins to the Greek word *kryos* for frost or icy cold. It collectively describes the portions of the Earth's surface where water is in a solid form and includes sea ice, lake ice, river ice, snow cover, glaciers, ice caps and ice sheets, and frozen ground (which includes permafrost). The cryosphere is an integral part of the global climate system with important linkages and feedbacks generated through its influence on surface energy and moisture fluxes, clouds, precipitation, hydrology, and atmospheric and oceanic circulation (Figure 6.1). Through these feedback processes, the cryosphere plays a significant role in global climate and in climate model response to global change.

A further concern of global warming is that melting ice sheets will cause sea level to rise. Even modest increases in sea level (~ 30 cm) are significant for coastal communities and coastal engineering (Asrar and Dozier 1994), and the economic and societal implications are immense. The Intergovernmental Panel on Climate Change (IPCC) expects that projected climate warming will lead to rapid and pronounced reductions in seasonal snow cover, permafrost, and glaciers (Fitzharris 1996). While the time scale for this response is uncertain, these changes can be expected to have widespread and significant impacts since the cryosphere is closely intertwined in the natural and economic fabric of many midlatitude and northern countries. In Canada, for example, most regions experience at least 3 months of snow cover each winter; nearly all navigable waters (with the exception of the west coast) are affected by an ice cover for some period during the winter; more than half of the country is underlain by continuous or discontinuous permafrost; and Canadian terrestrial ice masses constitute the most exten-

FIGURE 6.1



Schematic diagram outlining a number of the important interactions between the cryosphere and other major components of the global climate system. ***(G. Flato.)***

sive permanent ice cover in the Northern Hemisphere outside of Greenland (Goodison and Brown 1997).

Because of the sensitivity of the cryosphere to temperature changes, accurate information on the rate and magnitude of changes in cryospheric elements is essential for policy and decision making, particularly over the high latitudes of the Northern Hemisphere, where climate warming is projected to be the greatest. Global climate models (GCMs) do not yet provide accurate simulations of the current climate over the Arctic (Bromwich et al. 1994). Significant improvements in the representation of cryospheric processes and in the understanding of cryosphere-climate linkages and feedbacks are required to reduce the uncertainties in high-latitude climate simulations. Satellite data are indispensable for this task because of their unique combination of capabilities such as repeat coverage over data-sparse areas, all-weather sensing, and the ability to derive information on important surface geophysical properties. Satellites also have the ability to obtain information on horizontal and vertical displacements and other changes, using repeat coverage and specialized techniques such as Synthetic Aperture Radar (SAR) interferometry.

The importance of the cryosphere is clearly reflected in the Earth Observing System (EOS) program. For example, EOS includes a dedicated cryospheric data archiving center, launch of new sensors and algorithm development activities for improved monitoring of snow, ice, and glaciers, and several Interdisciplinary Science Investigation (IDS) activities undertaking research to improve understanding and modeling of cryospheric processes, cryospheric variability, and cryosphere-climate interactions. The following sections provide background information on the important characteristics of the main components of the cryosphere, before moving on to look at the scientific rationale for cryospheric research and monitoring activities within EOS (Section 6.2). Section 6.3 looks at the measurements needed to answer the key science questions, and summarizes the cryospheric data measurement activities of EOS and associated international science programs. The chapter concludes (Section 6.4) with a summary of important EOS contributions to improved understanding and monitoring of the cryosphere.

6.1.2 Components of the cryosphere

Frozen water occurs on the Earth's surface primarily as snow cover, freshwater ice in lakes and rivers, sea ice, glaciers, ice sheets, and frozen ground and permafrost (perennially-frozen ground). The residence time of water in each of these cryospheric sub-systems varies widely. Snow cover and freshwater ice are essentially seasonal, and most sea ice, except for ice in the central Arctic, lasts only a few years if it is not seasonal. A given water particle in glaciers, ice sheets, or ground ice, however, may remain frozen for 10-100,000 years or longer, and deep ice in parts of East Antarctica may have an age approaching 1 million years. The concept of residence time (flux/storage) is important for the climate system. Water with short residence times participates in the fast-response regime of the climate system (atmosphere, upper-ocean layers, and land surface) that determines the amplitude and regional patterns of climate change. Long-residence-time components (e.g., ice sheets and the deep ocean) act to modulate and introduce delays into the transient response (Chahine 1992). However, the possibility of abrupt changes in the slow-response components of the climate system cannot be overlooked.

The majority of the world's ice volume is in Antarctica (Table 6.1), principally in the East Antarctic ice sheet. In terms of areal extent, however, Northern Hemisphere winter snow and ice extent comprise the largest area, amounting to an average 23% of hemispheric surface area in January. The large areal extent and the important climatic roles of snow and ice, related to their unique physical properties, indicate that the ability to observe and model snow- and ice-cover extent, thickness, and physical properties (radiative and thermal properties) is of particular significance for climate research.

There are several fundamental physical properties of snow and ice that modulate energy exchanges between the surface and the atmosphere. The most important properties are the surface reflectance (albedo), the ability to transfer heat (thermal diffusivity), and the ability to change state (latent heat). These physical properties, together with surface roughness, emissivity, and dielectric characteristics, have important implications for observing snow and ice from space. For example, surface roughness is often the dominant factor determining the strength of radar backscatter (Hall 1996). Physical properties such as crystal structure, density, and liquid-water content are important factors affecting the transfers of heat and water and the scattering of microwave energy.

The surface reflectance of incoming solar radiation is important for the surface energy balance (SEB). It is the ratio of reflected to incident solar radiation, commonly referred to as albedo. Climatologists are primarily interested in albedo integrated over the shortwave portion of the electromagnetic spectrum (~0.3 to 3.5 μm), which coincides with the main solar energy input. Typically, albedo values for non-melting snow-covered surfaces are high (~80-90%) except in the case of forests (see textbox on pg. 266). The higher albedos for snow

TABLE 6.1

COMPONENT	AREA (10^6 KM^2)	ICE VOLUME (10^6 KM^3)	SEA LEVEL EQUIVALENT[a] (M)
Land Ice			
East Antarctica[b]	9.9	25.9	64.8
West Antarctica	2.3	3.4	8.5
Greenland	1.7	3.0	7.6
Small Ice Caps and Mountain Glaciers	0.68	0.18	0.5
Permafrost (excluding Antarctica)			
Continuous	7.60	0.03	0.08
Discontinuous	1.73	0.07	0.18
Sea Ice			
Northern Hemisphere[c]			
Late March	14.0	0.05	
Early September	6.0	0.02	
Southern Hemisphere[d]			
September	15.0	0.02	
February	2.0	0.002	
Land Snow Cover[e]			
Northern Hemisphere			
Late January	46.5	0.002	
Late August	3.9		
Southern Hemisphere			
Late July	0.85		
Early May	0.07		

Volumetric and areal extent of major components of the cryosphere.

a 400,000 km^3 of ice is equivalent to 1 m global sea level.
b Grounded ice sheet, excluding peripheral, floating ice shelves (which do not affect sea level). The shelves have a total area of 0.62 x 10^6 km^2 and a volume of 0.79 x 10^6 km^3 (Drewry 1982).
c Actual ice areas, excluding open water. Ice extent ranges between approximately 9.3 and 15.7 x 10^6 km^2.
d Actual ice area excluding open water (Gloersen et al. 1993). Ice extent ranges between approximately 3.8 and 18.8 x 10^6 km^2. Southern Hemisphere ice is mostly seasonal and generally much thinner than Arctic ice.
e Snow cover includes that on land ice, but excludes snow-covered sea ice (Robinson et al. 1993).

and ice cause rapid shifts in surface reflectivity in autumn and spring in high latitudes, but the overall climatic significance of this increase is spatially and temporally modulated by cloud cover. (Planetary albedo is determined principally by cloud cover, and by the small amount of total solar radiation received in high latitudes during winter months.) Summer and autumn are times of high-average cloudiness over the Arctic Ocean so the albedo feedback associated with the large seasonal changes in sea-ice extent is greatly reduced. Groisman et al. (1994a) observed that snow cover exhibited the greatest influence on the Earth radiative balance in the spring (April to May) period when incoming solar radiation was greatest over snow-covered areas.

The thermal properties of cryospheric elements also have important climatic consequences. Snow and ice have much lower thermal diffusivities than air (see text box). Thermal diffusivity is a measure of the speed at which temperature waves can penetrate a substance. As shown in the box, snow and ice are many orders of magnitude less efficient at diffusing heat than air. Snow cover insulates the ground surface, and sea ice insulates the

Typical Ranges for Surface Albedo:	
Fresh, dry snow	0.80 to 0.90
Melting ice/snow	0.25 to 0.80
Melting sea ice with puddles	0.30 to 0.40
Snow-covered forest	0.25 to 0.40
Snow-free vegetation/soil	0.10 to 0.30
Water (high solar elevation)	0.05 to 0.10

underlying ocean, decoupling the surface-atmosphere interface with respect to both heat and moisture fluxes. The flux of moisture from a water surface is eliminated by even a thin skin of ice, whereas the flux of heat through thin ice continues to be substantial until it attains a thickness in excess of 30 to 40 cm. However, even a small amount of snow on top of the ice will dramatically reduce the heat flux and slow down the rate of ice growth. The insulating effect of snow also has major implications for the hydrological cycle. In non-permafrost regions, the insulating effect of snow is such that only near-surface ground freezes and deep water drainage is uninterrupted (Lynch-Stieglitz 1994).

While snow and ice act to insulate the surface from large energy losses in winter, they also act to retard warming in the spring and summer because of the large amount of energy required to melt ice (the latent heat of fusion, 3.34×10^5 J kg^{-1} at 0$^{\circ C}$). However, the strong static stability of the atmosphere over areas of extensive snow or ice tends to confine the immediate cooling effect to a relatively shallow layer, so that associated atmospheric anomalies are usually short-lived and local to regional in scale (Cohen and Rind 1991). In some areas of the world such as Eurasia, however, the cooling associated with a heavy snowpack and moist spring soils is known to play a role in modulating the summer monsoon circulation (e.g., Vernekar et al. 1995). Gutzler and Preston (1997) recently presented evidence for a similar snow-summer circulation feedback over the southwestern United States.

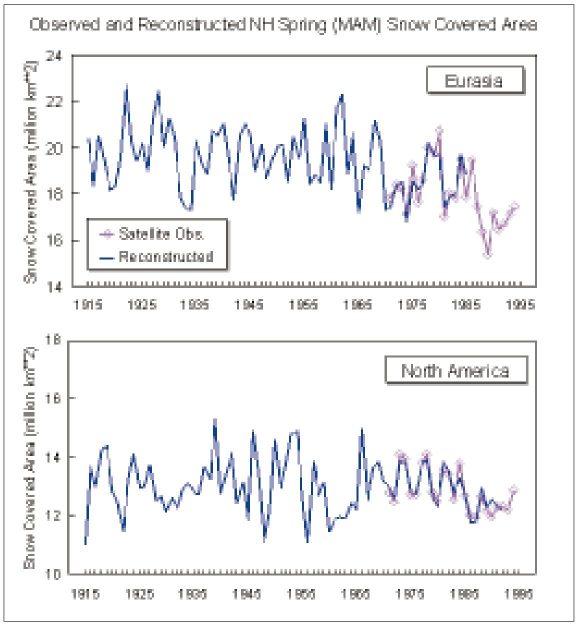
Typical Thermal Diffusivities: *(after Oke 1987)*	
	$m^2\ s^{-1} \times 10^{-6}$
Fresh snow	0.10
Old snow	0.40
Pure ice (0° C)	1.16
Air (still 10° C)	21.50
Air (turbulent)	~10^7

The role of snow cover in modulating the monsoon is just one example of a short-term cryosphere-climate feedback involving the land surface and the atmosphere. From Figure 6.1 it can be seen that there are numerous cryosphere-climate feedbacks in the global climate system. These operate over a wide range of spatial and temporal scales from local seasonal cooling of air temperatures to hemispheric-scale variations in ice sheets over time-scales of thousands of years. The feedback mechanisms involved are often complex and incompletely understood. For example, Curry et al. (1995) showed that the so-called “simple” sea ice-albedo feedback involved complex interactions with lead fraction, melt ponds, ice thickness, snow cover, and sea-ice extent. One of the main goals of EOS cryospheric research is to further the development of methods and models to observe and better understand interannual variations in cryospheric elements and their interactions with the global climate system. Further discussion of cryospheric variability and cryosphere-climate interactions is provided in the following subsections, which describe the key characteristics of the main components of the cryosphere, and in Section 6.2.1.

6.1.2.1 Snow

Snow cover has the largest areal extent of any component of the cryosphere, with a mean maximum areal extent of approximately 47 million km^2. Most of the Earth’s snow-covered area (SCA) is located in the Northern Hemisphere, and temporal variability is dominated by the seasonal cycle; Northern Hemisphere snow-cover extent ranges from 46.5 million km^2 in January to 3.8 million km^2 in August (Robinson et al. 1993). North American winter SCA has exhibited an increasing trend over much of this century (Brown and Goodison 1996; Hughes et al. 1996) largely in response to an increase in precipitation (Groisman and Easterling 1994). However, the available satellite data show that the hemispheric winter snow cover has exhibited little interannual variability over the 1972-1996 period, with a coefficient of variation (COV=s.d./mean) for January Northern Hemisphere snow cover of $<$ 0.04. According to Groisman et al. (1994a) Northern Hemisphere spring snow cover should exhibit a decreasing trend to explain an observed increase in Northern Hemisphere spring air temperatures this century. Preliminary estimates (Figure 6.2) of SCA from historical and reconstructed in situ snow-cover data suggest this is the case for Eurasia, but not for North America, where spring snow cover has remained close to current levels over most

Figure 6.2



Reconstructed SCA from PC analysis of in situ snow-cover data over North America and Eurasia *(Brown, 1997).*

of this century (Brown 1997). Because of the close relationship observed between hemispheric air temperature and snow-cover extent over the period of satellite data (IPCC 1996), there is considerable interest in monitoring Northern Hemisphere snow-cover extent for detecting and monitoring climate change.

Snow cover is an extremely important storage component in the water balance, especially seasonal snowpacks in mountainous areas of the world. Though limited in extent, seasonal snowpacks in the Earth's mountain ranges account for the major source of the runoff for stream flow and groundwater recharge over wide areas of the midlatitudes. For example, over 85% of the annual runoff from the Colorado River basin originates as snowmelt. Snowmelt runoff from the Earth's mountains fills the rivers and recharges the aquifers that over a billion people depend on for their water resources. Further, over 40% of the world's protected areas are in mountains, attesting to their value both as unique ecosystems needing protection and as recreation areas for humans. Climate warming is expected to result in major changes to the partitioning of snow and rainfall, and to the timing of snowmelt, which will have important implications for water use and management. These changes also involve potentially important decadal and longer time-scale feedbacks to the climate system through temporal and spatial changes in soil moisture and runoff to the oceans (Walsh 1995). Freshwater fluxes from the snow cover into the marine environment may be important, as the total flux is probably of the same magnitude as desalinated ridging and rubble areas of sea ice (Prinsenberg 1988). In addition, there is an associated pulse of precipitated pollutants which accumulate over the arctic winter in snowfall and are released into the ocean upon ablation of the sea-ice cover.

Variations in snowfall have also been shown to be important in the climate of polar regions since snow on sea ice affects both the thermodynamic and radiative characteristics of the ocean-sea ice-atmosphere interface. Brown and Cote (1992) demonstrated that the insulating effect of snowfall was the most important factor in determining the interannual variability of fast-ice thickness in the Canadian Arctic. In the coupled climate-sea-ice model of Ledley (1991, 1993), the albedo effect was found to dominate, and the effect of additional snowfall was to delay ice break-up and produce a cooling over the Arctic Basin. However, recent model sensitivity studies (Flato and Brown 1996; Harder 1997) suggest that the insulating effect is more important. In reality, it is extremely difficult to assess the baseline range of snow-thickness effects on the thermodynamic coupling between snow and sea ice because of a nearly complete lack of spatially-distributed estimates of snow cover on sea ice. EOS researchers are in the process of developing microwave algorithms for estimating snow-water equivalent over sea ice, which will address this problem (Cavalieri and Comiso 1997).

The snow-cover control of shortwave (SW) energy exchange at the surface also affects the availability of sub-ice photosynthetically-active radiation (PAR); an essential determinant of lower trophic level productivity. The evolution of the marine ecosystem is such that small perturbations in the availability of PAR have dramatic impacts on the initial production of epontic (sub-ice) algae (Welch 1992).

6.1.2.2 Sea ice

Sea ice covers much of the polar oceans and forms by freezing of sea water. Satellite data since the early 1970s reveal considerable seasonal, regional, and interannual variability in the sea-ice covers of both hemispheres. Seasonally, sea-ice extent in the Southern Hemisphere varies by a factor of 5, from a minimum of 3-4 million km^2 in February to a maximum of 17-20 million km^2 in September (Zwally et al. 1983; Gloersen et al. 1992). The seasonal variation is much less in the Northern Hemisphere where the confined nature and high latitudes of the Arctic Ocean result in a much larger perennial ice cover, and the surrounding land limits the equatorward extent of wintertime ice. Thus, the seasonal variability in Northern Hemisphere ice extent varies by only a factor of 2, from a minimum of 7-9 million km^2 in September to a maximum of 14-16 million km^2 in March (Parkinson et al. 1987; Gloersen et al. 1992).

The ice cover exhibits much greater regional-scale interannual variability than it does hemispherical. For instance, in the region of the Seas of Okhotsk and Japan, maximum ice extent decreased from 1.3 million km^2 in 1983 to 0.85 million km^2 in 1984, a decrease of 35%, before rebounding the following year to 1.2 million km^2 (Gloersen et al. 1992). The regional fluctuations in both hemispheres are such that for any several-year period of the satellite record some regions exhibit decreasing ice coverage while others exhibit increasing ice cover (Parkinson 1995). The overall trend indicated in the passive microwave record from 1978 through mid-1995 shows that the extent of Arctic sea ice is decreasing 2.7% per decade (Johannessen et al. 1995). Subsequent work with the satellite passive-microwave data indicates that from late October 1978 through the end of 1996 the extent of Arctic sea ice decreased by 2.9% per decade while the extent of Antarctic sea ice increased by 1.3% per decade (Cavalieri et al. 1997).

While small in terms of area, leads and polynyas are very important components of the sea-ice regime because of the dominant role they play in exchanges of heat and moisture to the atmosphere during the polar winter. Thermodynamically, polynyas have been observed to affect climate both at the local and regional scales (Steffen and Ohumura 1985), and the continuous ice production in polynyas is an important contribution to thermohaline circulation (Martin et al. 1992). Radiatively, polynyas become “oases” of biological production because of the significant absorption of PAR within the water column. This stimulates primary production, thereby creating a local abundance in biomass.

6.1.2.3 Lake ice and river ice

Ice forms on rivers and lakes in response to seasonal cooling. The sizes of the ice bodies involved are too small to exert other than localized climatic effects. However, the freeze-up/break-up processes respond to large-scale and local weather factors, such that considerable interannual variability exists in the dates of appearance and disappearance of the ice. Long series of lake-ice observations can serve as a proxy climate record, and the monitoring of freeze-up and break-up trends may provide a convenient integrated and seasonally-specific index of climatic perturbations. Information on river-ice conditions is less useful as a climatic proxy because ice formation is strongly dependent on river-flow regime, which is affected by precipitation, snow melt, and watershed runoff as well as being subject to human interference that directly modifies channel flow, or that indirectly affects the runoff via land-use practices. For various reasons including the small scale of river-ice processes, there is currently no specific river-ice research component within EOS. This may change in the future, but, for the moment, river ice is excluded from further discussion in this chapter.

Lake freeze-up depends on the heat storage in the lake and therefore on its depth, the rate and temperature of any inflow, and water-air energy fluxes. Information on lake depth is often unavailable, although some indication of the depth of shallow lakes in the Arctic can be obtained from airborne radar imagery during late winter (Sellman et al. 1975) and spaceborne optical imagery during summer (Duguay and Lafleur 1997). The timing of breakup is modified by snow depth on the ice as well as by ice thickness and freshwater inflow.

The appearance and disappearance of lake ice are readily observed in the visible band (Maslanik and Barry 1987; Wynne et al. 1996) and microwave sensors are particularly well-suited to monitoring important lake-ice characteristics such as freeze-up, break-up, freeze-to-bottom, and surface-melt onset (Hall 1993). Even small lakes (of a few square kilometers) can be monitored because freeze-up usually occurs after the development of a snow cover, and break-up after the disappearance of snow. Because ice freeze-up and break-up dates exhibit a close relationship to surface air temperatures at local and regional scales, monitoring of these readily observable events by satellite data offers the potential to monitor surface air-temperature anomalies in major data-sparse areas of the globe (Barry and Maslanik 1993). Recent closure or automation of many high-latitude stations increases the desirability of lake-ice monitoring via satellite in order to provide adequate spatial coverage of temperature trends in subarctic and Arctic regions. There are two major research activities being carried out in the Cryospheric System (CRYSYS) IDS to develop methods for using passive (Walker and Davey 1993) and active (Duguay and Lafleur 1997) microwave satellite data to monitor important lake-ice parameters.

Useful empirical relationships have been developed between temperature indices and the date of ice formation. In general, freeze-up and break-up correlate well with air temperature or freezing-degree days as used by Williams (1971). In general, empirical studies suggest an air temperature sensitivity of ice freeze-up/break-up of ±5 to 10 days per ±1°C seasonal change in air temperature (Palecki and Barry 1986). Thawing-degree days are less satisfactory indicators of breakup, especially where river inflow and wind are important factors or where there is considerable year-to-year variability in the depth of snow pack on the ice. Physical models of lake-water temperature and ice-cover growth and ablation (e.g., Fang and Stefan 1996) are a more-satisfactory way to proceed since these provide useful information on key processes and sensitivity to climate change, and are less restricted by the need to develop locally-calibrated statistical relationships.

6.1.2.4 Frozen ground and permafrost

Permafrost (perennially frozen ground) may occur where mean annual air temperatures (MAAT) are less than -1 or -2˚C and is generally continuous where MAAT are less than -7˚C. In addition, its extent and thickness are affected by ground moisture content, vegetation cover, winter snow depth, and aspect. The global extent of permafrost is still not completely known, but it underlies approximately 20% of Northern Hemisphere land areas. Thicknesses exceed 600 m along the Arctic coast of northeastern Siberia and Alaska, but, toward the margins, permafrost becomes thinner and horizontally discontinuous. The marginal zones will be more immediately subject to any melting caused by a warming trend. Most of the presently existing permafrost formed during previous colder conditions and is therefore relic. However, permafrost may form under present-day polar climates where glaciers retreat or land emergence exposes unfrozen ground. Washburn (1973) concluded that most continuous permafrost is in balance with the present climate at its upper surface, but changes at the base depend on the present climate and geothermal heat flow; in contrast, most discontinuous permafrost is probably unstable or “in such delicate equilibrium that the slightest climatic or surface change will have drastic disequilibrium effects” (Washburn 1973, p. 48).

Under warming conditions, the increasing depth of the summer active layer has significant impacts on the hydrologic and geomorphic regimes. Thawing and retreat of permafrost have been reported in the upper Mackenzie Valley and along the southern margin of its occurrence in Manitoba, but such observations are not readily quantified and generalized. Based on average latitudinal gradients of air temperature, an average northward displacement of the southern permafrost boundary by 50-to-150 km could be expected, under equilibrium conditions, for a 1˚C warming.

Only a fraction of the permafrost zone consists of actual ground ice. The remainder (dry permafrost) is simply soil or rock at subfreezing temperatures. The ice volume is generally greatest in the uppermost permafrost layers and mainly comprises pore and segregated ice in Earth material. Measurements of bore-hole temperatures in permafrost can be used as indicators of net changes in temperature regime. Gold and Lachenbruch (1973) infer a 2-4˚C warming over 75 to 100 years at Cape Thompson, Alaska, where the upper 25% of the 400-m thick permafrost is unstable with respect to an equilibrium profile of temperature with depth (for the present mean annual surface temperature of -5˚C). Maritime influences may have biased this estimate, however. At Prudhoe Bay similar data imply a 1.8˚C warming over the last 100 years (Lachenbruch et al. 1982). Further complications may be introduced by changes in snow-cover depths and the natural or artificial disturbance of the surface vegetation.

The potential rates of permafrost thawing have been established by Osterkamp (1984) to be two centuries or less for 25-meter-thick permafrost in the discontinuous zone of interior Alaska, assuming warming from -0.4 to 0˚C in 3-4 years, followed by a further 2.6˚C rise. Although the response of permafrost (depth) to temperature change is typically a very slow process (Osterkamp 1984; Koster 1993), there is ample evidence for the fact that the active-layer thickness quickly responds to a temperature change (Kane et al. 1991). Whether, under a warming or cooling scenario, global climate change will have a significant effect on the duration of frost-free periods in both regions with seasonally- and perennially-frozen ground.

6.1.2.5 Glaciers and ice sheets

Ice sheets are the greatest potential source of global freshwater, holding approximately 77% of the global total. This corresponds to 80 m of world sea-level equivalent, with Antarctica accounting for 90% of this. Greenland accounts for most of the remaining 10%, with other ice bodies and glaciers accounting for less than 0.5% (Table 6.1). Because of their size in relation to annual rates of snow accumulation and melt, the residence time of water in ice sheets can extend to 100,000 or 1 million years. Consequently, any climatic perturbations produce slow responses, occurring over glacial and interglacial periods. Valley glaciers respond rapidly to climatic fluctuations with typical response times of 10-50 years (Oerlemans 1994). However, the response of individual glaciers may be asynchronous to the same climatic forcing because of differences in glacier length, elevation, slope, and speed of motion. Oerlemans (1994) provided evidence of coherent global retreat in glaciers which could be explained by a linear warming trend of 0.66˚C per 100 years.

While glacier variations are likely to have minimal effects upon global climate, their recession may have contributed one third to one half of the observed 20th-century rise in sea level (Meier 1984; IPCC 1996). Furthermore, it is extremely likely that such extensive glacier recession as is currently observed in the Western Cordillera of North America (Pelto 1996), where runoff from glacierized basins is used for irrigation and hydropower, involves significant hydrological and ecosystem impacts. Effective water-resource planning and impact mitigation in such areas depends upon developing a sophisticated knowledge of the status of glacier ice and the mechanisms that cause it to change. Furthermore, a clear understanding of the mechanisms at work is crucial to interpreting the global-change signals that are contained in the time series of glacier mass-balance records.

Combined mass-balance estimates of the large ice sheets carry an uncertainty of about 20%. Studies based on estimated snowfall and mass output tend to indicate that the ice sheets are near balance or taking some water out of the oceans (Bentley and Giovinetto 1991). Marine-based studies (Jacobs et al. 1992) suggest sea-level rise from the Antarctic or rapid ice-shelf basal melting. Some authors (Paterson 1993; Alley 1997) have suggested that the difference between the observed rate of sea-level rise (roughly 2 mm y^{-1}) and the explained rate of sea-level rise from melting of mountain glaciers, thermal expansion of the ocean, etc. (roughly 1 mm y^{-1} or less) is similar to the modeled imbalance in the Antarctic (roughly 1 mm y^{-1} of sea-level rise; Huybrechts 1990), suggesting a contribution of sea-level rise from the Antarctic.

Relationships between global climate and changes in ice extent are complex. The mass balance of land-based glaciers and ice sheets is determined by the accumulation of snow, mostly in winter, and warm-season ablation due primarily to net radiation and turbulent heat fluxes to melting ice and snow from warm-air advection (Munro 1990; Paterson 1993; Van den Broeke 1996). However, most of Antarctica never experiences surface melting (Van den Broeke and Bintanja 1995). Where ice masses terminate in the ocean, iceberg calving is the major contributor to mass loss. In this situation, the ice margin may extend out into deep water as a floating ice shelf, such as that in the Ross Sea. Despite the possibility that global warming could result in losses to the Greenland ice sheet being offset by gains to the Antarctic ice sheet (Ohmura et al. 1996), there is major concern about the possibility of a West Antarctic ice-sheet collapse. The West Antarctic ice sheet is grounded on bedrock below sea level, and its collapse has the potential of raising the world sea level 6-7 m over a few hundred years.

Most of the discharge of the West Antarctic ice sheet is via the five major ice streams (faster flowing ice) entering the Ross Ice Shelf, the Rutford Ice Stream entering Ronne-Filchner shelf of the Weddell Sea, and the Pine Island Glacier entering the Amundsen Ice Shelf. Opinions differ as to the present mass balance of these systems (Bentley 1983, 1985), principally because of the limited data. The West Antarctic ice sheet is stable so long as the Ross Ice Shelf is constrained by drag along its lateral boundaries and pinned by local grounding.

Over the last few decades, a number of major programs have significantly improved understanding of the Greenland ice sheet (e.g., Expédition Glaciologique Internationale au Groenland (EGIG), Camp Century, Dye 3, Greenland Ice Sheet Program [GISP]). Reeh (1989) concluded that the Greenland ice sheet as a whole is close to a balanced state with slight thinning of some marginal sectors likely being compensated for by a slight thickening in the central area. Koerner (1989) provided evidence that the Greenland ice sheet experienced extensive melt during the last interglacial 100,000 years ago, which suggests that the current ice sheet is not a relic from a previously colder climate. This was confirmed in an ice-sheet modeling study when, after removal, the ice sheet re-formed on bare bedrock under present or slightly warmer climatic conditions (Letréguilly et al. 1991). The model experiments also revealed that the ice sheet was relatively stable, and that it took temperature rises of at least 6˚C for the ice sheet to disappear completely. A recent doubled-CO_2 simulation of the ice sheet with the Global Environmental and Ecological Simulation of Interactive Systems (GENESIS) GCM (Thompson and Pollard 1997) indicated that the net annual mass balance decreased from +0.13 to -0.12 m a^{-1}. However, there are still major uncertainties involved in such modeling exercises, particularly in the area of how atmospheric circulation and precipitation will change. Ice-sheet-model sensitivity to accumulation parameterization was highlighted in Fabre et al. (1994), where, for a temperature increase of 5˚C, one accumulation parameterization yielded only slight margin retreat while the other resulted in the complete collapse of the ice sheet.

Satellites are playing an increasingly important role in furthering understanding of the Greenland ice sheet mass balance by providing accurate surface-elevation data, ice-flow velocity data (Joughin et al. 1995; Rignot et al. 1995), and surface-ablation information. For example, Abdalati and Steffen (1995) used passive microwave data to observe an increasing trend in ablation area of 3.4% per year since 1978. This trend was temporarily interrupted by the aerosol loading from the eruption of Mt. Pinatubo in 1991, which underscores the sensitivity of the mass balance to perturbations in incoming solar radiation.

6.2 Major scientific questions

As outlined in IPCC (1996, p. 46) the most urgent scientific problems requiring attention are 1) determining the rate and magnitude of climate change and sea-level rise, 2) the detection and attribution of climate change, and 3) the regional patterns of climate change. Addressing these questions requires an improved understanding and ability to model the entire coupled atmosphere-ice-ocean climate system, as well as systematic observations of key climate variables such as snow and sea-ice extent. Improved representation of the cryosphere is vital given documented weaknesses in GCM climate simulations over high latitudes, the area expected to exhibit the greatest warming in response to increased levels of greenhouse gases. With regard to the cryosphere the key areas are:

- improved representation of cryospheric processes and cryosphere-climate interactions in climate and hydrological models, and

- improved ability to monitor and understand variability and change in important components of the cryosphere such as major ice sheets and hemispheric snow and sea-ice extent.

The following sections elaborate the current state of cryospheric science in these key areas and outline the ways in which EOS will contribute to improved understanding of the cryosphere.

6.2.1 Representation of cryospheric processes in climate and hydrological models

Previous evaluations of the performance of GCMs at simulating the present Arctic climate (e.g., Walsh and Crane 1992; Bromwich et al. 1994; McGinnis and Crane 1994) documented numerous shortcomings such as incorrect seasonal variation in cloud cover, excessive atmospheric moisture transport, poor simulation of atmospheric circulation patterns around Greenland, and an excessive dominance of winter-season circulation patterns. Bromwich et al. (1994) noted that the poor simulation of atmospheric circulation around Greenland is greatly improved with higher-resolution representations of Greenland topography by T42 coupled models. However, the problem of excessive northern-latitude precipitation persists in T42 models (Bromwich and Chen 1995). Battisti et al. (1997) concluded that most of the GCMs used to evaluate climate change have an artificially dampened natural variability in the Arctic, and that this was primarily linked to inadequate treatment of sea-ice processes (e.g., snow, albedo, heat storage, and conduction). Another cryospheric link to poor climate-model performance was highlighted by Tao et al. (1996) who found that a lack of vegetative masking of snow albedo in some Atmospheric Model Intercomparison Project (AMIP) models resulted in a 3.3˚C cold bias in spring air temperatures over northern Eurasia. A high priority for cryospheric research is to improve the representation of the cryosphere in hydrologic models and in global and regional climate models.

6.2.1.1 Snow cover

Snow cover in climate models

The realistic simulation of snow cover in climate models is essential for correct representation of the SEB (albedo, surface temperature, heat and moisture fluxes) ground temperatures, and hydrology. Snow cover has a number of important effects on the hydrological cycle. First, it acts as a major store for winter precipitation, which is subsequently released in the spring. Second, the insulating properties of snow cover are such that in all but permafrost regions, frost penetration into the soil is limited to surface layers, which allows deep water drainage to continue throughout the year (Lynch-Stieglitz 1994). In early GCMs (e.g., the National Center for Atmospheric Research [NCAR]-CCM0A), snow cover was specified from a monthly snow-cover climatology, and the only impact of snow cover was through changes in surface albedo. Studies of GCM performance with and without physically-based snow-cover models (Marshall et al. 1994, Lynch-Stieglitz 1994) have highlighted the need to include an adequate treatment of snow cover to model correctly the hydrological cycle. Lynch-Stieglitz (1994) concluded that the inability to correctly model snow cover cast doubt upon year-round model calculations of runoff.

The energy and mass-balance model of Anderson (1976) is probably one of the most complete snowpack models developed to date. However, this model is too computationally-intensive for GCMs, which have rather severe computational constraints. A review of current GCM snow-cover models (Verseghy 1991; Loth et al. 1993; Marshall et al. 1994; Lynch-Stieglitz 1994) suggests that the following processes must be properly represented in order to provide realistic simulations of the snowpack:

- Snow metamorphism or aging: Snow experiences major changes in physical properties (density, albedo, thermal conductivity) once it reaches the surface due to a number of processes such as melt/freezing, settling, compaction, and water vapor diffusion. These changes are time- and temperature-dependent and need to be taken into account in order to model correctly the energy balance of the snowpack. A number of empirically-based parameterizations are available which describe these processes, and the verification results appear to be good for the few case studies presented in the literature to date (Lynch-Stieglitz 1994).

- Ability to resolve vertical gradients: Both Loth et al. (1993) and Lynch-Stieglitz (1994) conclude that a multi-layer representation of the snowpack is essential to model the steep temperature gradient at the snow surface and to take account of depth-varying snow properties. Lynch-Stieglitz (1994) recommended that the depth of the surface layer be no greater than the thermal-damping depth of snow (~6-10 cm) to resolve diurnal fluctuations in heat content.

- Snow-rain separation: GCMs employ a variety of schemes for determining precipitation phase (see Table 2 in Randall et al. 1994). Loth et al. (1993) tested several rain/snow criteria and found that the modeled snow cover and runoff (for a midlatitudinal maritime climate) were highly sensitive to the particular criterion used.

Of the various GCM snowpack models published in the literature to date, the multi-layer model of Loth et al. (1993) is the most detailed and potentially the most realistic for 1-D snowpack simulations. However, Lynch-Stieglitz (1994) obtained excellent results with a simplified three-layer snow-cover model, which suggests that important snowpack processes can be captured without a great deal of vertical resolution. It should be noted that most GCM snowpack-model evaluations carried out to date have assessed performance at a single point; none have been validated systematically over a full range of snow-cover climate regimes such as those identified by Sturm et al. (1995). A general lack of suitable evaluation data is a major contributing factor. EOS products from the Moderate-Resolution Imaging Spectroradiometer (MODIS), such as surface temperature, albedo, and snow cover, as well as snow-water equivalent and depth from the EOS Advanced Microwave Scanning Radiometer (AMSR-E), will greatly assist this task.

A major challenge for GCM snow-cover modelers is inclusion of important physical processes, and topographic and land-cover factors affecting the spatial and temporal distribution of snow cover and its properties. For example, Pomeroy et al. (1997) demonstrated that winter precipitation alone was insufficient to calculate snow accumulation in Arctic regions, and that blowing-snow processes and landscape patterns were the key factors governing the spatial distribution of snow-water equivalent in winter. A recent validation of a point snowpack model highlighted the problem of ignoring wind redistribution processes (Yang et al. 1997). In this case, the model validation results were worse when corrected gauge-catch data were used as the performance standard, because the model did not account for blowing-snow losses. Stochastic approaches can be used to account for sub-grid-scale variations in vegetation (Woo and Steer 1986) and topography (Stieglitz et al. 1997; Walland and Simmonds 1996). In the latter case, the parameterization of sub-grid-scale topographic variability in a GCM snow sub-model yielded more-accurate simulations of seasonal variation in Northern Hemisphere SCA, and solved the characteristic GCM problem of retarded spring snow-cover retreat noted by Frei and Robinson (1995). GCMs also exhibit a general tendency to overestimate snow cover over the Tibetan Plateau and China, but this problem is probably due more to atmospheric model difficulties simulating temperature and precipitation in the lee of a major mountain barrier than to systematic errors in simulating snowpack processes.

The evidence suggests that the multi-layer-snow energy-balance models used in GCMs are able to capture the essential one-dimensional character (surface temperature, depth, snow-water equivalence, density, and snowmelt) of a midlatitudinal snowpack, and that current GCMs are able to provide, for the most part, reasonable simulations of seasonal and year-to-year variation in continental-scale SCA (Foster et al. 1996). However, many of the important processes affecting regional-to-local-scale variations in snow cover and snow properties are not taken into account in current GCM snow models. These shortcomings have important consequences for local-regional-scale hydrology and climate, particularly as the spatial resolution of GCMs increases. EOS satellites will play a major role in helping to address these problems by providing higher-quality land-cover data sets, and higher-resolution snow-water-equivalent observations from sensors such as AMSR-E for model validation and investigation of scaling issues.

Snow cover in hydrological models

The above discussion applies to coarser-scale GCM representations of snow cover. However, in order to predict watershed response to changes in climate or atmospheric pollution, one needs a sensitive, accurate modeling capability that adequately captures the major processes controlling snowmelt, runoff, and chemical changes in soil and lakes. This information is needed at a scale corresponding to the scale of variability in watershed geology, soils, and vegetation that is important in controlling these changes. While empirical snowmelt-runoff models have been useful for operational runoff-volume forecasts, they provide little information on the timing, rate, or magnitude of discharge, and they are inappropriate in situations outside the boundary conditions governing the development of the relevant empirical parameters. Thus, they may fail to predict water yield adequately in extreme or unusual years, and they cannot be reliably used in investigations examining snowmelt responses to climate variability and change. These problems, and the increasing importance of understanding intrabasin snowmelt dynamics for environmental analysis of such factors as basin ecology (Baron et al. 1993), water chemistry (Wolford et al. 1996), and hillslope erosion (Tarboton et al. 1991), have motivated the development of physically-based, spatially-distributed snowmelt models in recent years. Such models require information on the spatial distribution of snowpack water storage. But, mountain snowpacks are spatially heterogeneous, reflecting the influences of rugged topography on precipitation, wind redistribution of snow, and surface energy fluxes during the accumulation season (Elder et al. 1991).

Snowmelt drives the hydrology in alpine basins and snow-cover changes over the melt season are of primary interest to water managers in the western U.S. A distributed snowmelt model consists of two general components: a model for calculating melt at a single point (given a set of prescribed snowpack and meteorological conditions), and a method of developing the requisite snowpack and meteorological data for all points within the basin. Snowmelt-modeling efforts require several steps necessary to couple basin-wide energy-balance snowmelt models with remote sensing and flow routing. The model TOPORAD (Dozier and Frew 1990) uses information on watershed topography (i.e., a Digital Elevation Model [DEM]) to distribute radiation spatially. Simpler models are used to determine the spatial distribution of other energy-balance components. These radiation maps are then used as inputs to models that estimate the distribution of snow around the basin prior to snowmelt (i.e., initial conditions for melt), and as inputs to the point snowmelt model. To predict the biogeochemical response of catchments, models such as AHM (Wolford et al. 1996) use calculations of water generated at the snow surface coupled with a model of elution of water and chemical species from the snowpack, and routes that water through the basin to estimate, or predict, streamflow and its chemical concentrations.

One of the main obstacles to physically-based modeling is the accumulation of the necessary meteorological and snow-cover data to run, calibrate, and validate such models. For example, basin discharge has frequently been used as the sole physical criterion of model calibration and performance assessment for conceptual snowmelt models. But as basin discharge is an integrated response to melt and runoff, it does not discriminate among the various effects of a multiplicity of data inputs that drive physical models. Distributed snow-cover data are required if one is to properly assess model performance in such complex terrain (Bloschl et al. 1991a; Bloschl et al. 1991b). No widely suitable method yet exists to directly map snow-water equivalence in rugged mountain regions. However, the higher-resolution snow products generated through EOS will help address this issue. Estimates of snow-covered area based on remote-sensing data can significantly improve the performance of even simple snowmelt models in alpine terrain (Kite 1991; Armstrong and Hardman 1991). Rango (1993) reviewed the progress that has been made incorporating remote-sensing data into regional hydrologic models of snowmelt runoff. For operational purposes, empirical approaches using remote-sensing data to estimate snow-covered area and snow-depth networks to estimate snow-water equivalence are continuing to improve (Martinec and Rango 1991; Martinec et al. 1991).

Both visible and near-infrared, and passive microwave remote-sensing data are being used to develop estimates of snow-covered area for operational forecasting of snowmelt. Development of accurate snow-cover information for areas with steep, variable topography requires higher-resolution data than are currently available from operational remote-sensing instruments. Models using the higher-resolution satellite data expected during the next decade show good results with test data acquired from aircraft platforms. Determination of other snowpack properties in alpine areas, such as grain size and albedo (from visible/infrared) and snow-water equivalence (from active microwave) are topics of continuing research. Progress in both algorithm development and testing with field data sets shows that obtaining these properties is achievable. The volume-integrating capability of microwave remote sensing has received much attention, because it offers the possibility of remote determination of whole snowpack properties. However, the lack of multipolarized SAR will limit the ability to reliably estimate snow-water equivalence in alpine areas.

6.2.1.2 Sea ice

Model simulations of the climatic impact of increasing greenhouse gases typically show enhanced warming at high latitudes, largely as a result of positive feedbacks involving sea ice. A concern is that sea ice has been treated rather simply in climate models. In particular, sea ice in GCMs has generally been approximated as a motionless thermodynamic slab. Inclusion of sea-ice dynamics—advection and deformation—is necessary to account for the freshwater provided by melting sea ice advected from the Arctic into the Greenland and Norwegian Seas, and to reproduce dynamic/thermodynamic feedback effects involving open-water "leads." A further motivation for improving the representation of sea ice in climate models is that sea-ice changes have a major impact on the availability of moisture and the mass balance of major ice sheets, especially Antarctica (Rind et al. 1995). Of the 16 GCMs summarized in the recent IPCC report (Gates et al. 1996), only six included any form of ice motion. Of those six, only three included a real ice dynamics scheme (based on solution of the sea-ice momentum equation); the remaining three simply transported ice with the ocean-surface current.

Experiments with stand-alone sea-ice models (e.g., Hibler 1984; Lemke et al. 1990) have shown that thermodynamic-only models are more sensitive to changes in thermal forcing than those that include dynamics. The reason is, at least in part, the constant formation of open-water leads, which dominate overall ice growth. Recent GCM experiments by Pollard and Thompson (1994) have shown that inclusion of sea-ice dynamics produced more-realistic ice extent and reduced the model's globally-averaged CO_2-induced warming, with the effect most pronounced around Antarctica.

Including a more-realistic dynamical formulation is the principal improvement currently being made to the sea-ice component of many GCMs. The two approaches most widely used are the "viscous-plastic" model of Hibler (1979) and the simpler "cavitating fluid" model described by Flato and Hibler (1992). Both schemes produce similar large-scale thickness build-up and transport, but they differ in the details of their treatment of internal ice stresses. An important scientific question is "what is the optimal sea-ice dynamics model for use in climate simulations?" This question is being addressed in part through the World Climate Research Program (WCRP) Arctic Climate System Study (ACSYS) project (see Section 6.3.2.1) in the form of the Sea Ice Model Intercomparison Project (SIMIP) (Lemke et al. 1996). SIMIP involves comparison of a hierarchy of sea-ice models to one another and to available observations, using the same forcing and boundary conditions, in order to illustrate the advantages and disadvantages of various parameterizations and their suitability for use in global climate modeling. Such an effort requires accurate specification of atmospheric and oceanic forcing, along with observations of ice thickness, concentration, velocity, and its spatial derivatives (i.e., deformation). Observations of these same quantities are also required for studies of the ice mass balance and its variability as outlined in Section 6.3.1.2.

Sea-ice thermodynamic schemes used in climate models also tend to be rather crude. In particular, the surface-albedo parameterization (Barry 1996) and the calculation of heat conduction and internal heat storage are typically highly idealized. Surface albedo is highly variable owing to the mixture of open water and ice, some of which may be covered by snow (which itself has a range of albedos). In spring, this situation is further complicated by the presence of melt ponds that arise from surface melt and subsequently evolve in both depth and areal extent due to absorption of radiation and drainage. Storage and release of heat within the ice cover affects the timing of growth and melt. Because of the short summer-melt season, and the large solar radiative fluxes in summer, this effect can have a substantial impact on the ice mass balance. Recent studies by Bitz et al. (1996) and Battisti et al. (1997) have investigated the role of ice thermodynamics in a single-column climate model. Their results indicate that relatively modest changes in thermodynamic parameterizations may have a disproportionate effect on simulated ice-thickness variability, and that this in turn has a major impact on the simulated climate variability. Development and validation of sea-ice thermodynamic parameterizations suitable for use in global climate simulations is therefore a further scientific challenge for EOS.

6.2.1.3 Lake ice

Lakes are important components of many ecosystems, particularly in northern boreal and tundra environments, where lakes and standing water occupy a significant fraction (~25%) of the total land cover. Lakes influence local energy and water exchanges, and the freezing and thawing of lake ice has important consequences for physical, chemical, biological, and hydrological processes (Heron and Woo 1994). Physical models have been developed that are able to simulate lake-water temperatures and ice cover over long periods, e.g., Fang and Stefan (1996). However, most lakes are not resolved in current GCMs, and, according to Arpe et al. (1997), the neglect of frozen water stored in lakes and rivers has led to incorrect simulation of the seasonal cycle of discharge from the Mackenzie River, and to an incorrect supply of freshwater to the Arctic Ocean in coupled models. With increasing resolution of Numerical Weather Prediction (NWP) and regional climate models, there is growing interest in explicit representation of lake processes. For example, the

CLASS Land Surface Process (LSP) model will incorporate a lake-model component in 1998 (D. Verseghy, personal communication 1997).

A number of EOS initiatives are being carried out to understand the physical processes and structural factors affecting the microwave signatures of ice-covered lakes (e.g., Walker and Davey 1993; Hall et al. 1994; Jeffries et al. 1994; Duguay and Lafleur 1997). These initiatives are providing important validation data for lake-ice-model development and testing (conventional shore-based ice observations are inadequate for this purpose). In addition, the coupling of numerical ice-growth models and SAR data has provided important ancillary information such as the depth distribution of northern lakes.

6.2.1.4 Frozen ground and permafrost

Frozen ground plays a significant role in the terrestrial portion of the hydrological cycle because it restricts moisture exchanges between surface water and deep ground water (Prowse 1990). The occurrence of frozen ground and permafrost is, therefore, an important factor controlling drainage and the areal and spatial distribution of wetlands (Rouse et al. 1997). Permafrost is also important for modeling snow cover as the cold sub-surface layer increases the amount of energy required to melt the snowpack and delays melt (Marsh 1991).

A recent example highlighting the insights that can be gained through application of 1-D models was provided by Zhang et al. (1996), who used a 1-D finite-difference heat-transfer model to investigate the sensitivity of the ground thermal regime to variations in the depth-hoar content of the overlying snow cover. Depth hoar (or sugar snow) is a relatively low-density (~150-250 kg m^{-3}) layer of snow with large rounded grains that forms near the ground surface in response to steep temperature gradients. The depth-hoar fraction can be over 50% in tundra snow, and can represent up to 80% of the snowpack for taiga snow (Sturm and Johnson 1991). Zhang et al. (1996) showed that changes in the depth-hoar fraction from 0 to 60% could increase daily ground-surface temperatures by 12.8˚C and mean annual surface temperatures by 5.5˚C. This result underscores the importance of including realistic simulations of vertical snow-density variations in climate and hydrological models.

To date, few hydrological or climate models include frozen-ground processes in any detail, which results in poor performance at simulating the hydrological cycle in northern environments. The inclusion of frozen soil in hydrological models is a major objective of the Canadian Global Energy and Water Cycle Experiment (GEWEX) program. In LSP models with multi-layer treatments of soil heat and water flow (e.g., CLASS, Verseghy 1991), soil freezing and thawing can be simulated theoretically. However, accurate simulation of permafrost processes such as active-layer development will require LSP schemes to include additional soil layers and better understanding of heat and water flow in organic soils (Woo, personal communication, 1997). There are still LSP schemes in use that ignore soil freezing completely (Verseghy, personal communication, 1997).

EOS permafrost research activities are focusing on the development of techniques for mapping the spatial distribution of permafrost using remotely-sensed and ancillary data, and the development of algorithms for detecting changes in active-layer depth and terrain features characteristic of degrading permafrost such as ground-ice slumps and thaw lakes. These activities will generate high-resolution information on frozen-ground extent for input and validation of LSP schemes and distributed hydrological models.

6.2.1.5 Glaciers and ice sheets

Of the roughly 7 mm y^{-1} of sea-level equivalent that is added to the surfaces of the great ice sheets, less than 1 mm y^{-1} is returned to the sea as meltwater running off the upper surfaces. The great majority of return to the oceans is by ice flow, either as icebergs broken off the coastal regions or as melting from the undersides of floating extensions of the ice sheets called ice shelves. Furthermore, as discussed below, much larger changes in ice flow are possible that are likely to affect surface-balance estimates. Should large changes occur, rapid ice-sheet thinning and sea-level rise could result, but not rapid ice-sheet thickening or sea-level drop. This inherent asymmetry of ice behavior, slow growth but rapid decay, has dominated the ice-age cycles of the last million years (Imbrie et al. 1993) and figures prominently in the view of the future.

Ice outflow is driven by gravity acting on regions with sloping ice surfaces. Gravitational-driving stress increases with surface slope and the ice thickness. Ice responds to this stress by deforming internally and by moving over its substrate (Paterson 1994). The rate of internal deformation increases with temperature and with the cube of stress (Paterson 1994), and depends on the physical properties produced by the history of deformation (Alley 1992). First principles indicate small uncertainties in deformation-rate estimates; tuned models closely match observations. Motion over the bed is much more complex and may account for only a small fraction of total velocity, but in regions called ice streams, which have exceptionally lubricated beds, ice velocity may be 100 times higher than for adjacent ice with similar stress. Basal velocity is achieved by sliding over the substrate, or by deformation within the substrate (Alley and Whillans 1991).

Internal deformation responds slowly to perturbations. Increased snowfall causes the ice to thicken. This increases the driving stress and the thickness through which the ice flows, both of which increase the ice flux until it matches the increased accumulation and a new steady state is reached. Increased surface temperature eventually warms and softens the ice, which increases ice outflow, thinning the ice and reducing the driving stress until a new steady state is reached. However, because most deformation occurs in deep ice, heat must penetrate near the bed before it causes significant changes. Marginal retreat steepens the ice sheet, which increases the gravitational stress and the ice flow, forcing a wave of thinning that propagates inland. For existing ice sheets, adjustments to changes in snowfall and marginal position require a few thousand years, and the response to temperature changes is somewhat longer. Assessment of internal deformations indicates that of the modern ice sheets, Greenland responds most rapidly and East Antarctica most slowly (Whillans 1981; Alley and Whillans 1984). The Greenland ice sheet has almost completely adjusted to the end of the last ice age, West Antarctica has adjusted for the most part, although further changes can be expected, and East Antarctica has completed only initial responses. Internal deformations appear to be unstable for large ice masses because rapid deformation creates heat, which softens the ice for further deformation; however, this is probably not a dominant process (Clarke et al. 1977; Huybrechts and Oerlemans 1988).

Basal processes are much more difficult to model and predict. Bifurcations and instabilities certainly exist. For example, a glacier frozen to its bed does not move rapidly over that bed, but a thawed-bed glacier can (MacAyeal 1993a, 1993b). Where the bed is thawed, frictional heating increases water supply, thus increasing basal velocity up to some maximum value. Beyond that point basal velocity may decrease with further increases to water supply as the flow system transforms from a distributed one that lubricates ice motion, to a channelized one that does not (Weertman 1972). Most ice motion may occur through deformation of subglacial sediments (Alley and Whillans 1991), but large uncertainties still exist in the understanding and modeling of such processes.

Of great interest is the likelihood that the extreme variations in ice-sheet behavior from place to place also occur at one place over time. Mountain glaciers with thawed, or partly thawed beds are known to experience large, rapid switches in velocity, called surges, that are primarily associated with basal processes (Paterson 1994). Observations in West Antarctica have shown exceptionally large time evolution, with some regions thickening and others thinning (Alley and Whillans 1991). Glacial-geological evidence shows major changes in ice-marginal positions over very short times for former ice sheets (Clark 1994), and evidence from marine sediments (the Heinrich events; Broecker 1994) indicates large ice surges from Hudson Bay and other ice sheets during the most recent North American glaciation. However, work on such flow instabilities remains in the explanatory rather than the predictive stage.

The most widely discussed instability is that of marine ice sheets (Mercer 1968; Weertman 1974). In the simplest model, an ice sheet with a bed below sea level that deepens towards the center of the ice sheet is inherently unstable. Either its grounding line, which separates ice that is too thick to float free of the bed from floating ice, must advance to the edge of the continental shelf, or it must retreat to the center of the ice sheet, collapsing the ice and raising sea level. Some stability can be provided if the floating extension is not free to spread in the ocean because it runs aground on islands or is constrained in an embayment. The modern West Antarctic ice sheet has a bed which deepens towards the center around most of the coast, and it has constrained ice shelves around much of its coast. In the disaster scenario, warmed waters circulating beneath the ice shelves would melt them partially or completely, reducing their constraint on the grounded ice and triggering a collapse that would raise sea level meters in centuries. The possibility of such a disaster remains; the likelihood is debated (IPCC 1990, 1996; Huybrechts 1990).

Other mechanisms of instability include thawing of frozen regions (Clarke et al. 1984), and loss or narrowing of interstream ridges that now provide much restraint on fast-moving ice streams (Whillans et al. 1993). As for marine instability, present knowledge does not allow exclusion or prediction of such behavior.

The prevailing view of ice-sheet response to climatic change (IPCC 1990, 1996) holds that a coming global warming will enhance melting of the Greenland ice sheet and enhance snowfall on the Antarctic ice sheet, with little net effect on sea-level change. If Greenland ice sheet melt is offset by growth in the Antarctic ice sheet, then the contribution of small glaciers to sea-level rise is not merely significant, as indicated by Meier (1984); it has the potential to be the principal cryospheric agent for sea-level change (Ohmura et al. 1996). Thus the impacts of small-glacier retreat in mountainous areas extend well beyond implications for local water supply.

But predictions about the great ice sheets are disseminated with large, asymmetric error bars that allow the small, but significant possibility of ice-sheet collapse in West Antarctica. A “Delphic oracle” assessment of the possibility of ice-sheet collapse leads to a somewhat higher probability of rapid sea-level rise than suggested in the IPCC documents (Titus and Narayanan 1995). Whole-ice-sheet models which parameterize accumulation to increase with temperature (Letréguilly et al. 1991; Huybrechts 1993) typically yield slow, steady ice-sheet adjustment occurring over millennia, or longer. The changes in snowfall and surface melting then dominate the contribution to sea-level change over the century—or centuries-long “planning horizons” adopted. Modeled warming of more than 6˚C removes the Greenland ice sheet entirely over 10,000 years (Letréguilly et al. 1991). Over roughly a 10,000-year period, a model warming of 8-10˚C will remove the West Antarctic ice sheet, while increasing this amount to approximately 20° C suffices to remove the East Antarctic ice sheet (Huybrechts 1993). These models, however, typically lack the “fast physics” of basal velocity (occurring through sliding or bed deformation) that is strongly sensitive to small perturbations. One model of the West Antarctic ice sheet that included such “fast physics,” when integrated through ten 100,000-year ice-age cycles, frequently produced rapid volume changes, and complete ice-sheet collapse during three of the ten cycles (MacAyeal 1992).

Major uncertainties in projecting future behavior include:

- significant uncertainty in current ice-sheet and glacier mass-balance estimates, particularly as they apply to areal representation;
- questions of how future climate change will be manifested regionally and amplified in polar and alpine regions;
- the difficulty of predicting how atmospheric circulation and precipitation will change in response to global warming;
- the uncertainties of how the “fast physics” of ice flow works, and how to incorporate it into models; and
- the questions of whether future perturbations (perhaps through ice-shelf basal melting) or past perturbations (perhaps through the ongoing penetration of the post-ice-age warmth to the ice-sheet beds) can trigger rapid changes in ice flow and sea level.

Glaciers and ice sheets in climate models

Theoretically, fully-coupled GCMs have the potential to simulate the main processes (precipitation, accumulation, ablation) affecting the mass balance of glaciers and ice sheets, and to capture important feedbacks (such as changes in oceanic and atmospheric circulation) that would accompany a significant reduction in land-ice volume. Until recently, one of the main obstacles to simulating ice-sheet mass balances in GCMs was that of scale (Thompson and Pollard 1997). Accurate representation of the topography of ice sheets (e.g., steep edge slopes) is critical for mass-balance calculations because it affects a host of parameters and processes such as surface temperature, precipitation, and topographic interactions with atmospheric circulation. Recent higher-resolution GCMs (~200-km grid spacing) have been able to provide much-more-realistic simulations of the Antarctic and Greenland ice-sheet mass balances (e.g., Ohmura et al. 1996; Thompson and Pollard 1997). The other main obstacle to simulating ice sheets in GCMs is that, to date, they have not explicitly included ice-flow dynamics and processes such as re-freezing of meltwater. This is less of a shortcoming over short time scales, but is important for examining issues such as sea-level change beyond the end of the next century when ice-sheet dynamic response is significant (Huybrechts et al. 1991). Early attempts to couple GCMs with dynamical ice-sheet models (e.g., Verbitsky and Saltzman 1995) were hampered by the coarse resolution of GCMs, which adversely affected precipitation simulations over the major ice sheets. The inclusion of parameterizations for elevation effects on surface meteorology and for the refreezing of meltwater in the GENESIS GCM (Thompson and Pollard 1997) was found to produce much more realistic simulations of ice-sheet mass balance.

The IPCC (1996) report indicates four major knowledge gaps that need to be addressed in order for climate models to provide better estimates of glacier and ice-sheet mass balance and contributions to sea-level change:

1) Inclusion of the processes and feedbacks linking meteorology to mass balance and dynamic response.

2) Extension of glacier modeling and process research to a broader spectrum of glacier environments, in particular the larger glaciers mountains of Alaska, central Asia, and the ice caps of Patagonia and the Arctic.

3) Quantification of the process of meltwater refreezing.

4) Increased understanding of the process of iceberg calving.

EOS satellites will play a major role in helping to reduce these uncertainties: The Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) will provide the capability for glacier mass-balance monitoring at a global scale (Global Land Ice Monitoring from Space [GLIMS]); Earth Science Enterprise (ESE) altimetry data are providing detailed surface topography information for input to atmospheric and ice-sheet models (Ice, Clouds, and Land Elevation Satellite [ICESat]); SAR interferometric methods are being applied to data from the European Remote-Sensing Satellite-1/2 (ERS-1/2) and the Radar Satellite (Radarsat) to generate detailed observations of surface topography and ice-flow rates (e.g., Vachon et al. 1996), and passive-microwave and visible-satellite data from EOS satellites will provide critical information for monitoring and understanding key processes such as surface melt.

Glaciers and ice sheets in hydrological models

One of the major links between glaciers and ice sheets and the global climate system is through runoff. The ability to correctly model glacier-runoff processes is, therefore, essential for realistic simulation of future mass-balance and sea-level changes, and the thermohaline circulation. The importance of runoff is further underscored through the coupling of glacier hydrology and glacier flow dynamics (Iken 1981; Iken and Bindschadler 1986; Kamb 1987; Alley 1989; Harbor et al. 1997). This is mediated by the status of subglacial drainage, which is thermomechanically controlled. Where subglacial drainage occurs, the key questions to address are the configuration of the drainage system (channelized/distributed), how it evolves on various time scales, and exactly where the drainage pathways are located. Answers to these questions hold the key to understanding seasonal and subseasonal variations in glacier velocity, surge-type behavior, the formation of ice streams, and the detailed distribution of velocity in glacier cross sections.

Insofar as glacier response to climate forcing involves an immediate mass-balance response and a longer-term dynamic response, glacier hydrology may be important for the dynamic response (Kamb and Engelhardt 1991; Arnold and Sharp 1992; Alley 1996). Its role could be linked to changes in glacier thermal regime resulting from the combined effects of changing glacier geometry and atmospheric boundary conditions, changes in the length of the summer melt season, changes in runoff volume, changes in runoff regime (e.g., amplitude of diurnal cycles changing in response to thinning or thickening of snow cover, eradication of firn reservoirs), or changes in the distribution of subglacial drainage pathways (linked to changing glacier geometry). All of these could alter the existence, longevity, character, and distribution of subglacial drainage systems, and the nature of the water-storage regime within them, with implications for ice-flow dynamics.

An improved understanding of glacier hydrology is also needed with respect to glacier hazards. Outburst floods from supraglacial, ice marginal, and englacial/subglacial sources are a concern in high-mountain and volcanically-active areas, and there is a need to document possible source reservoirs (location and size, and frequency of flood release). The existence of very large lakes beneath ice sheets raises the possibility of long-interval, high-magnitude events that might be significant for global sea level and for ocean, circulation, and climate.

Improved understanding and modeling of glacier hydrology is also crucially important from a water-resources perspective. Glaciers are an important source of water for human and livestock consumption, irrigation, and hydro-power in high mountain regions. This is especially true at certain critical times of the year, such as the growing season, when water demand peaks and glacier runoff accounts for a significant fraction of the available water supply. For example, during August in the Bow River, Alberta, Young et al. (1996) determined that glacier ice wastage contributed almost one quarter of the total flow during low-flow years. Variations in glacier extent affect the magnitude of this resource in complex ways. It is important in this regard to understand that times of unusually high meltwater supply may also be times of accelerated ice-resource loss, and that effective water-supply planning must be done with the expected long-term status of the resource clearly in view.

Glacier and ice-sheet runoff are also important from an ecological perspective (e.g., habitat sustainability, wildlife management) through influences on stream-water turbidity and temperature (Brugman et al. 1997). Other water-quality-related ecology issues include the effects of supraglacial runoff. This runoff may be isolated from contact with soils which can buffer acid runoff, and as a result could produce severe acid shock in meltwater streams (Johannessen and Henriksen 1978).

6.2.2 Cryosphere-climate linkages and feedbacks

Reducing the uncertainties in projections of rates and regional patterns of climate change requires improved representation of climate processes in models, especially feedbacks associated with water vapor, clouds, oceans, ice and snow, and land-surface/atmosphere interactions (IPCC 1996). The most obvious cryosphere-climate feedbacks involve modifications to the SEB (albedo and insulating effects) and to moisture exchanges with the overlying atmosphere. There are also less obvious, but important indirect linkages between the cryosphere and the climate system that operate through the movement and storage of water in the freshwater cycle. Freshwater storage in the cryosphere occurs over a range of time scales as indicated in Section 6.1, and is therefore involved in climate variability over a similar range of time scales. In recent years, a growing body of evidence has accumulated suggesting that the climate of the North Atlantic is highly sensitive to variations in freshwater input, and that the climatic response can be sudden and dramatic (Walsh and Chapman 1990; Weaver and Hughes 1994).

Cryosphere-climate feedbacks can be classified into two main categories: feedbacks working through the freshwater cycle, and feedbacks working through the SEB. These are discussed in more detail below.

6.2.2.1 Freshwater cycle

The cryosphere is intimately connected to the freshwater cycle through the storage and transport of freshwater in solid form. Freshwater storage in the cryosphere occurs over a range of time scales and, therefore, is involved in climate variability over a similar range of time scales.

On short time scales (seasonal, annual), the accumulation and melting of snow dominates the hydrological cycle of many alpine and high-latitude drainage basins. In populated areas near glacier-fed river systems, such as the Bow, the Columbia, the Fraser, and the North Saskatchewan, the amount and timing of snowmelt-induced spring runoff has direct human consequences in terms of hydroelectric power generation, water supply, and flooding. Furthermore, valley glaciers respond rapidly to climatic fluctuations, such as those induced by increasing amounts of atmospheric carbon dioxide (Oerlemans 1986), so significant changes in length and volume may occur on the time scale of a few decades.

On annual-to-decadal time scales, sea ice can store freshwater and transport it from one region to another as ice drifts under the influence of winds and currents. This occurs because as sea ice freezes it expels most of the salt, leaving the ice relatively fresh; when ice forms in one location, is transported, and melts in another, the ocean experiences an imbalance in surface salt flux. This process is particularly important in the Northern Hemisphere as ice is transported out of the Arctic into the North Atlantic where it melts, thus forming a stabilizing freshwater layer at the surface. The freshwater layer may moderate deep convection in the North Atlantic and thereby affect the global “conveyor belt” circulation, which is important for transporting oceanic heat northward and sequestering carbon dioxide in the deep ocean. Variations in this export of sea ice may excite variability in oceanic deep convection and global thermohaline ocean circulation (Halekinen 1993).

On still-longer time scales, large glaciers and ice sheets can play an even more dramatic role, storing and releasing freshwater in amounts that may substantially raise or lower global sea level, thus posing problems for coastal settlements. Because of their vast size, climatic perturbations produce slow responses in these ice masses, on the time scale of glacial-interglacial periods. Recent modeling work (Chen et al. 1997) has highlighted the important role of topographically-related feedbacks between major ice sheets and atmospheric circulation. They presented evidence for a possible feedback between topographically-induced lee cyclogenesis and the mass balance of a large ice sheet, which has significant implications for warmer-world simulations of the Greenland ice sheet.

There are also important feedbacks between permafrost and the climate system through the hydrological cycle. Because the hydraulic conductivity of permafrost is significantly lower than unfrozen soil, this limits ground-water flow and is an important factor controlling drainage patterns and the spatial distribution of wetlands in northern ecosystems (Rouse et al. 1997). Permafrost is also a significant store of CO_2 and methane, which are released when ground thaws; and ground ice represents an important freshwater store for contribution to sea-level rise.

A major scientific challenge is to quantify the exchanges of freshwater between the hydrosphere and cryosphere, and to understand the role of such exchanges in climate variability and change. EOS will contribute to meeting this challenge through an improved ability to observe and simulate the mass budget of sea ice, glaciers, and ice sheets.

6.2.2.2 SEB

Perhaps the most direct and dramatic effect of the cryosphere on the climate system is its role in modifying the SEB, over both land and ocean. The two principal mechanisms involving ice and snow cover (Randall et al. 1994) are insulation of land and ocean surfaces from the atmosphere, or outgoing longwave radiation (OLR) feedback, and enhancement of the surface albedo, or shortwave feedback.

The SW feedback comes about through the classic positive feedback between temperature and albedo, where an increase in air temperature enhances melt, thus reducing ice and snow cover, which then decreases albedo, thus leading to greater absorbed energy and warmer temperatures. This feedback is an important issue in climate modeling insofar as it amplifies errors in the parameterization of the processes involved. It is also an important reason that GCMs predict enhanced warming in the polar regions as a result of increasing greenhouse gases. Recent climate-model results indicate that about 37% of a simulated global temperature increase caused by carbon dioxide doubling was the result of direct or indirect sea-ice feedbacks (Rind et al. 1995). In reality, the ice-albedo feedback is considerably more complex than described above. Curry et al. (1995) showed that the sea-ice-albedo feedback involves interactions with lead fraction, melt ponds, ice-thickness distribution, snow-cover and sea-ice extent, and that all the processes should be correctly parameterized in order for a climate model to yield the correct sensitivity to external forcing.

The ice-albedo feedback is inextricably linked with cloud-radiation feedbacks (Shine and Crane 1984; Curry et al. 1996), particularly over snow cover, where increases in cloud cover may increase the atmospheric emissivity sufficiently to offset the positive albedo feedback. The interaction of clouds and radiation with summertime melting of snow and sea ice is considered to be an important area of scientific uncertainty in understanding clouds and radiation in the Arctic (Curry et al. 1996). Over land the presence or absence of snow alters the surface albedo and hence the surface-radiation balance. This effect is much larger over bare soil or grasslands than over forested areas, where the effect of snow albedo is reduced by standing vegetation (see list of typical albedo values in Section 6.1.2).

The low thermal conductivity of snow slows the winter heat loss from the surface, which results in a lower OLR to the atmosphere. An ice cover also exerts a major influence on the exchange of sensible and latent heat with the atmosphere, allowing less heat loss from water as the ice cover thickens. Open water within the polar ice pack, which may be long, narrow leads, or irregular openings between floes, occupies only a small fraction of the ice-covered area (typically less than 10%), but dominates the SEB (Maykut 1978). This open water also contributes to the albedo-temperature feedback responsible for much of the polar amplification of greenhouse-gas warming obtained by climate models. Some estimates of the sensitivity of a climate model to open-water fraction are provided by Flato and Ramsden (1997) and references therein.

Snow and ice also affect the energy balance through the latent heat required to melt ice. The largest SEB feedbacks occur during the spring period, when incoming solar radiation to snow and ice cover is increasing, but heat gain at the ground is hindered by both the high albedo and the latent heat of fusion required to melt the cover. There are also numerous indirect feedbacks operating through the hydrological cycle (e.g., soil moisture and clouds) although these can still be considered as SW and OLR feedbacks. A good example of this is the role of snow in a monsoon-type circulation. One view (e.g., Barnett et al. 1989; Yang et al. 1997) is that a heavy spring snowpack reduces land-sea contrasts (i.e., higher albedo and soil moisture → cooler surface temperatures), which weakens the summertime land-surface heating driving the monsoon. As outlined in Meehl (1994) however, increased soil moisture also provides a moisture source for increased precipitation, so that soil moisture has two competing monsoon effects through increased evaporation and cooler surface temperatures. The insulating effect of snow on sea-ice growth has been shown to be an important climate feedback through its role in modulating sea-ice volume (Harder 1997).

6.2.2.3 Observed and modeled feedbacks

Numerous empirical studies have been carried out to quantify the effect of snow cover on temperature and large-scale circulation anomalies (see reviews by Cohen and Rind [1991] and Leathers and Robinson [1993]). These studies reveal that snow cover is associated with local temperature decreases in the range of 1˚-6˚C, and below-normal geopotential heights. Leathers and Robinson (1993) showed that snow cover could promote significant temperature perturbations well away from the area of anomalous snow cover through large-scale modification of air masses. As noted by Cohen and Rind (1991), a major difficulty with empirically-based studies is to separate the thermodynamic effect of snow cover from the dynamical influence of the regime which produced the snow cover in the first place. Is a winter with positive snow-cover anomalies colder than normal because of the snow cover, or is there more snow cover because of the colder temperatures? This separation is possible with GCMs where experiments can be run that isolate the effect of snow cover on climate. For example, Cohen and Rind (1991) applied the Goddard Institute for Space Studies (GISS) GCM to this problem and found that positive NH snow-cover anomalies in March caused only short-term local decreases in surface temperature. They proposed a negative feedback mechanism to explain these results, whereby the increased atmospheric stability caused by the snow cover acted to suppress the latent and sensible heat flux away from the surface, resulting in a gain in net heating and enhanced removal of the snow-cover anomaly.

Results from a single GCM, however, are insufficient for rendering a convincing picture of the snow-climate feedback. Evaluation of snow-climate feedbacks in several GCMs (Cess et al. 1991; Randall et al. 1994) revealed that the feedback differed markedly between models, ranging from a weak negative, to a strong positive feedback. One reason for the divergence in the results was that the snow feedback was observed to be associated with a number of complex effects caused by cloud interactions. In one experiment, for example (Cohen and Rind 1991), changes in cloud cover tended to offset much of the impact due to changes in snow cover. They found that although the change in surface albedo at 51˚ N between the high- and low-snow-cover runs was more than 20%, the absolute difference in planetary albedo at 51˚ N was only 2% due to the masking influence of cloud cover. In light of the acknowledged weakness of GCMs, with respect to representing clouds and cloud-climate feedbacks (IPCC 1990), GCMs may not yet adequately capture the full complexities of snow-climate feedbacks.

Satellites provide the capability to directly measure cryosphere-climate feedbacks through observations of the global extent of snow and ice cover as well as the main components of the Earth radiation budget. For example, Groisman et al. (1994a) used satellite-derived snow cover and Earth Radiation Budget Experiment (ERBE) data to obtain direct estimates of the snow-cover feedback on the radiation balance at the top of the atmosphere. Their results showed significant spatial and seasonal variations in the magnitude and sign of the feedback. In the fall and winter, the OLR-reduction effect of snow cover dominated, resulting in a positive radiation feedback. For clear-sky conditions, the cooling albedo effect only began to dominate the warming influence from reduced OLR in February and reached a maximum during the spring (March-May). The zones of maximum snow-cover feedback were located over Siberia and the eastern Arctic region of Canada. Groisman et al. (1994a) estimated that the impact of snow cover on Northern Hemisphere extratropical latitudes between high (1979) and low (1990) snow-cover years was 0.9 W m^{-2}, which corresponds to an increase of about 0.5˚C in annual surface-air temperature, or about half the observed temperature change between the two years.

An important scientific challenge is to improve the ability to simulate the important geophysical properties and processes controlling energy exchanges in the cryosphere, e.g., melt-pond formation, leads, surface-cloud radiative exchanges, snow density, ice-thickness distribution. EOS observations will play a major role in this process through provision of data for validating process models and through the direct observation of feedback processes.

6.2.3 Cryospheric variability and change

The ability to observe the natural variability of key cryospheric parameters such as sea-ice volume, snow and ice mass, and the area of snow and ice is critical for understanding climate-cryosphere feedbacks, for validation of climate models, and for climate-change detection. EOS satellites will be contributing to an enhanced ability to observe the cryosphere, which is a major objective of the WCRP Global Climate Observing System (GCOS). A summary of the GCOS requirements for cryospheric data is presented in Cihlar et al. (1997). For most cryospheric variables, however, the period of available satellite data is too short to provide insights into the decadal and longer time-scale variability known to be important in the climate system. This will require the application of techniques to combine in situ observations, physical models, and satellite data. The following sections outline current understanding and knowledge gaps in observing cryospheric variability and change.

6.2.3.1 Snow

Satellites are well suited for mapping global-scale variations in snow cover, and two important databases exist for investigations of spatial and temporal variability in hemispheric and global snow cover: the weekly National Oceanic and Atmospheric Administration (NOAA) visible satellite-based snow-cover analysis from 1972[1] (Robinson et al. 1993), and Scanning Multispectral Microwave Radiometer (SMMR) and Special Sensor Microwave/Imager (SSM/I) passive microwave brightness temperature data from 1978, which can be used to derive information on snow extent (Chang et al. 1987), depth (Foster et al. 1984), and snow-water equivalence (Goodison 1989; Chang et al. 1991). The spatial and temporal character of these two data sets is relatively coarse (190.5-km polar stereographic grid and 1/week for the NOAA data set, and ~25-km resolution and 1/day for the passive microwave data) but this is sufficient for monitoring key properties such as snow extent, dates of snow-cover onset and disappearance, and peak accumulation, which are known to be important indicators of change (Barry et al. 1995). The NOAA data set is known to have a number of shortcomings in areas with persistent cloud cover, low solar illumination, dense forest, patchy snow, and mountainous regions (Robinson et al. 1993). However, the most important characteristic of the

[1] Regular satellite monitoring of NH snow cover began in November 1966 (Dewey and Heim, 1982), but the sub-point resolution of the pre-1972 satellites was ~4.0 km compared to 1.0 km with the VHRR launched in 1972 (Robinson et al., 1993). A project to re-chart and digitize the pre-1972 data was recently completed at Rutgers U., which will extend the NOAA record back to 1966 (D. Robinson, *personal communication 10/30/97*).

data set for studying snow-cover variability is that the analysis methodology has been relatively stable for over 25 years (Basist et al. 1996). The passive microwave data are not limited by solar illumination or cloud cover, and offer a higher resolution. The problem is applying snow algorithms that give reliable results over a range of land-cover types and snow-cover conditions. Evaluations of hemispheric maps of SSM/I-derived snow-water equivalence (Foster et al. 1996) and snow extent (Basist et al. 1996) report systematic biases related to vegetation effects and melting snow. Some of these problems can be overcome by including a wet-snow indicator (e.g., Walker and Goodison 1993) and applying land-cover-specific algorithms following Goita et al. (1997). The development and validation of snow algorithms is a major EOS activity during the pre-launch period for MODIS (Hall et al. 1996) and AMSR-E (Chang and Rango 1996). An automated snow-cover analysis scheme based on SSM/I was implemented by the National Environmental Satellite Data and Information Service (NESDIS) in November 1997 but will be run in parallel with the manual analysis for a period of 18 months to determine possible impacts on the homogeneity of the existing database.

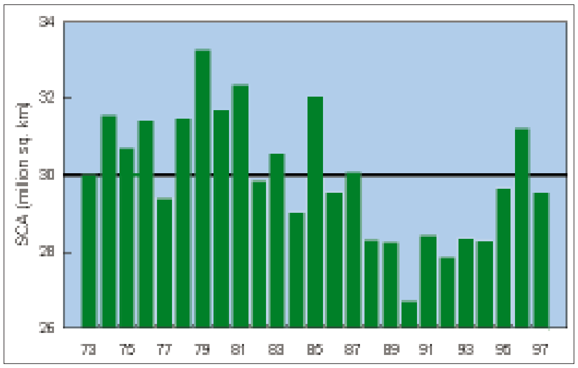
The existing satellite snow databases are currently too short to provide information on the natural variability in continental-scale snow cover needed for climate-change detection and for validating transient simulations of GCMs. Regular in-situ daily snow-depth observations are available in some regions of the world from as early as the late 1800s. However, there are few stations with complete records prior to the early 1900s. Historical snow-cover data sets suitable for studies of snow-cover variability are available for the United States (Easterling et al. 1997; Hughes and Robinson 1996), Canada (Brown and Goodison 1996) and the former Soviet Union (Fallot et al. 1997). Typical problems encountered when working with historical snow data include inconsistent observing practices, inhomogeneities, and bias in observing networks. In spite of this, it appears that useful information on continental-scale variability in snow cover can be obtained from in situ data sets, particularly for variables such as seasonal snow-cover duration, which can be approximated from daily temperature and precipitation data, and which exhibit high spatial coherence (Brown and Goodison 1996). For example, Brown (1997) applied principal-component analysis to observed and reconstructed in situ snow-cover-duration data to reconstruct spring SCA over North America and Eurasia from 1915 to 1985 (Figure 6.2). (See Section 6.1.2.1.) The reconstruction method was able to explain 81% of the springtime variance in satellite-observed SCA over North America and 67% over Eurasia. The results suggested spring SCA had decreased significantly over Eurasia this century, but there was no evidence of a similar long-term decrease in North America. These results are consistent with observed 20th century midlatitudinal spring air-temperature trends over the same period that show a statistically significant increase over Eurasia, but not over North America. Further efforts are needed to develop other methods and approaches for integrating in situ and satellite data to obtain additional independent estimates of historical variation in hemispheric snow cover.

Documenting the spatial and temporal variability of snow depth and snow water-equivalence poses more of a challenge than SCA or snow-cover duration. Snow-water-equivalence algorithms are still being developed and tested for forested environments (Goita et al. 1997), deep snowpacks (de Sève et al. 1997), and over sea ice (Markus and Cavalieri 1997). SAR offers potential for high-resolution mapping of wet or melting snow, which is important for operational monitoring of mountain snowpacks and for integration into runoff models (Hall 1996). EOS researchers are also investigating methods for extracting snow-water-equivalence information from dry and shallow snowpacks using Radarsat data (Bernier and Fortin 1998). Historical data sets of in situ snow-water-equivalence data are available over some regions of the world (e.g., the Former Soviet Union) from 1966-1990 (Armstrong and Krenke 1997) and over Canada from ~1960 (Braaten 1997). While spatially and temporally constrained, these data sets are important for model validation (e.g., Yang et al. 1997). The development and validation of passive microwave snow-water-equivalence algorithms is a key activity for understanding spatial and temporal variability in snow-water equivalence and is vital for snow-cover-validation activities being carried out under AMIP. The collection of detailed validation data sets is an important contribution to understanding spatial and temporal variability in snow cover, and for the correct representation of this variability in climate models.

Given the decadal and long-term variability known to be important in the climate system, it is misleading to present trends of hemispheric snow cover based on the short period of available satellite data, unless appropriately qualified. Groisman et al. (1994a) reported a 10% decrease in Northern Hemisphere annual snow cover over the 1972-1991 period using the NOAA weekly snow-cover data set which was quoted in the recent IPCC (1996) assessment of climate change. More-recent satellite data show that the rapid decrease in Northern Hemisphere spring SCA that characterized the 1980s appears to have reversed itself during the 1990s (Figure 6.3).

Longer historical snow-cover time series indicate that regional snow cover exhibits considerable variabil-

FIGURE 6.3



Annual variation in NH spring (MAM) SCA in million km² derived from NOAA weekly snow-cover analyses. The horizontal line is the mean spring SCA over the 1973-1997 period. ***(Data from NOAA Climate Prediction Center.)***

ity on time scales from years to decades (e.g., Koch and Rüdel 1990; Brown and Goodison 1996; Hughes and Robinson 1996; Fallot et al. 1997). Proxy data can also yield useful information on snow-cover variability. For example, Lavoie and Payette (1992) documented a 20th-century increase in snow cover over subarctic Québec of 0.4 m based on increases in the level of windblown snow abrasion on Black Spruce. Palaeo data on lake levels (Vance et al. 1992) and flood frequency (Knox 1993) suggest that snow cover and snowmelt over the continental interior of North America have likely undergone large and rapid changes in response to abrupt changes in temperature and precipitation that occurred ~5000 years ago (warm and dry), ~3,300 years ago (cool and wet), during the Medieval Warm Period (warm and dry), and again during the Little Ice Age (cool and wet). In conclusion, no single sensor or method alone can provide the information needed to observe spatial and temporal variability in snow cover at the scales required for climate-change monitoring and detection (Barry et al. 1995). Various sensor systems have advantages and disadvantages, and integration of satellite data, in situ observations, and physical models is needed to document the natural variability in the cryospheric system.

6.2.3.2 Sea ice

Changes in the energy balance, as might accompany climatic warming, would affect the ice-mass balance. In addition to possible climatic feedbacks, changes in ice thickness or areal coverage have direct consequences for maritime transportation, offshore oil and gas developments, and the marine ecosystem. Observations of the sea-ice mass balance are needed to quantify its variability so that significant changes can be identified, to understand the processes which contribute to variability and change, and to verify the representation of sea ice and its interactions in climate models.

Determining the mass or volume of sea ice from observations requires measurements of both areal coverage and thickness. There are two aspects to areal coverage: ice concentration and ice extent. Concentration is defined as the fraction of area, in a given location, covered by ice; the remainder being open water. Ice extent is defined as the area enclosed by the ice edge, which is often defined as the 15% concentration isoline. Rather good observations of ice concentration, and hence ice extent, have been available from satellite passive microwave radiometry since the 1970s. However, complications arise in interpreting the satellite data, a primary one being the matching of records from different instruments (Bjorgo et al. 1997). It thus becomes desirable for EOS to obtain long records from the same instrument and to have overlap periods with alternative records. Some alternative sources include aircraft and shipborne observations assembled primarily for operational ice analysis and forecasting. Despite problems of spatial coverage and quality, these observations allow assembly of longer time series, such as that produced by Walsh (1978) which, for the Arctic, extends back to 1900. AMSR-E will be the most important EOS instrument for continuing the satellite record of sea-ice concentration and extent.

So far there is no method available to measure ice thickness from satellites, and techniques like airborne electromagnetic surveys are still experimental. A recent development in satellite remote sensing designed to address this issue is the Radarsat Geophysical Processor System (RGPS) housed at the Alaska SAR facility in Fairbanks, Alaska. This system is designed to ingest SAR data of the entire Arctic Basin. These data will be georectified onto a Lagrangian model grid, and displacement statistics will be computed for each of the Lagrangian grids. This procedure will produce weekly ice thickness as a function of ice kinematic histories (openings and closings) throughout the annual cycle at the kilometer scale (Kwok et al. 1995). The first fields of ice-thickness data from RGPS will be available in the third quarter of 1997. At present, the only "operational" method of surveying ice thickness is via upward-looking sonar. Examples of submarine sonar compilations of ice draft are given by Bourke and Garrett (1987) and McLaren et al. (1992). It should be noted that ice thickness varies dramatically over scales as small as a few tens of meters, from open water to ridged ice, which may be greater than 20 m in thickness. Thickness observations and models of ice-thickness evolution are therefore couched in terms of the "ice-thickness distribution" (the probability density of ice thickness [e.g., Thorndike et al. 1975]). The mean of this thickness distribution is most relevant to the sea-ice mass balance, and model studies like that of Flato (1995) may be of use in developing a measurement strategy. A pressing scientific need is to promote access to, and analysis of, submarine thickness data, along with data from moored upward-looking sonar instruments, to complement EOS observations. Considerable progress has been made in this regard through joint efforts of Vice President Al Gore of the United States and Prime Minister Viktor Chernomyrdin of Russia. As a result of their initiative, considerable Arctic submarine data from both countries have been declassified and are being made available to the science community (Belt 1997).

In addition to simply observing the amount of sea-ice mass, it is also important to observe the principal fluxes that enter the mass balance. On a "basin scale", these include thermodynamic growth and melt, ridging, and transport. These fluxes, and the processes which determine them, are often the means by which the cryosphere interacts with other parts of the climate system. For example, the transport of sea ice out of the Arctic and into the Greenland Sea, where it melts, is a mechanism whereby the cryosphere affects water-mass-formation rates and ventilation of the deep ocean.

To summarize, a complete sea-ice mass budget requires observations of ice-covered area and thickness, the net SEB, and ice transport and deformation. In addition to mass balance, however, it is also important to quantify the spatial and temporal variability in the other characteristics of the ice cover that are significant in terms of ice-climate interaction (e.g., surface albedo, surface temperature, lead distribution). The distribution of open water within the pack and its variability through time can be measured using SAR data, while both ice albedo and surface temperature will be provided from MODIS.

6.2.3.3 Lake ice

Lake-ice freeze-up and break-up, particularly break-up, have been found to be useful indicators of regional climate change (Palecki and Barry 1986; Reycraft and Skinner 1993). Records of lake- or river-ice break-up spanning 100 or more years exist in a number of locations with some extending back to the 16th century (Kuusisto 1993). More dense networks of lake-break-up observations are available from ~1950 (Canada) and from the 1920s for Scandinavia.

The combination of physical lake models, in situ observations, and satellite data is required to document natural variability in lake-ice cover over large areas. The major advantage of satellite data is that they provide objective information on lake freeze-up/break-up over the entire lake surface, unlike shore-based observations. Satellite observations are therefore particularly valuable for validating/calibrating physical lake-ice models, which provide the link for combining satellite and in situ records of lake freeze-up/break-up. Satellite-derived lake-ice information can be obtained from a range of sensors including visible imaging systems such as the Advanced Very High Resolution Radiometer (AVHRR), passive microwave systems, and SAR. ICEMAP (using MODIS) will produce sea-ice-duration information for the larger inland lakes. The use of passive microwave data is limited to the larger lakes, but has the advantage of all-weather capability. SAR, however, may be useful for measuring freeze-up and break-up conditions over small lakes.

6.2.3.4 Frozen ground and permafrost

Since climate is the dominant factor influencing the continental distribution of permafrost, the monitoring of permafrost conditions and geographic extent should be useful for deriving information on high-latitude climate change (Leverington and Duguay 1996). Satellites are ideally suited for monitoring the vast, uninhabited areas underlain by permafrost. However, there are several characteristics of permafrost that pose serious challenges for the remote observation of permafrost extent and change. Permafrost conditions cannot be directly observed by orbiting sensors in the way that exposed cryosphere elements such as snow cover, sea-ice and lake-ice cover, or glacier extent can be. First, permafrost is a sub-surface feature, which means that extent and change have to be indirectly deduced from related micro-climate and surface-vegetation characteristics that can be detected with remote-sensing techniques (Hall and Martinec 1985). Second, permafrost responds to the ground heat flux, which is affected by a host of local factors (e.g., vegetation cover, snow cover, sub-surface thermal properties) in addition to external climatic factors, and the response involves significant time lags on the order of decades (Leverington and Duguay 1996).

A number of studies have been able to successfully discriminate frozen/unfrozen ground and the depth of the late-summer active layer with high-resolution multispectral imagery such as is obtained by Landsat Thematic Mapper (TM) (e.g., Morrissey et al. 1986; Peddle et al. 1994; Leverington and Duguay 1996; Leverington and Duguay 1997). This success, however, is closely linked to local training data. Leverington and Duguay (1997) found that classification performance dropped from 90% to 60% when a trained neural-net classification scheme was transferred to a similar area a few tens of kilometers away from the training site (the change in surficial deposits being the key factor behind the difference in permafrost-surface relations between the two study areas).

On the basis of their research, Leverington and Duguay (1996) concluded that remotely-sensed monitoring of permafrost has a number of obstacles to overcome before it can be used for large-scale climate monitoring over high latitudes. These include:

- the vertical resolution in estimates of depth to frozen ground is currently insufficient for detecting changes over time;
- land-cover permafrost correlative relationships may change over time;
- the significant time lag between changes in climate, and change in surface cover and permafrost (this time lag may be on the order of decades).

In spite of these problems, remotely-sensed classification of frozen ground remains a powerful tool for regional-scale permafrost research, environmental management, and hydrological modeling. With nearly 50% of Canada, 80% of Alaska, and roughly 25% of the continents underlain by permafrost, extensive ground surveys of permafrost conditions are precluded owing to expense, logistical difficulties, short field seasons, and time constraints. On the other hand, many numerical heat-transfer models have been used in understanding the detailed physics of permafrost-climate relationship. However, these models are impractical beyond the site scale because of the extremely limited database characterizing the microclimates of a broad range of vegetation and terrain conditions. Satellite remote sensing provides the means to spatially distribute the models (local to regional) and therefore more effectively evaluate the impact of climate change on permafrost.

High-resolution remote-sensing imagery is particularly useful for monitoring thermokarst features and the annual freeze-thaw cycle of the landscape. This type of imagery is particularly relevant in the context of climate change. As the greatest amounts of ground ice are usually found near the permafrost table, it is expected that thermokarst features could occur at the beginning of climatic warming within a period of only a few decades (Koster 1993). For example, optical (Landsat TM and Systeme pour l'Observation de la Terre MLA/PLA) and active microwave (Airborne Synthetic Aperture Radar [AIRSAR], ERS-1) data have been used successfully to map permafrost thaw features (e.g., retrogressive thaw slumps, active-layer detachment slides, and thaw lakes (Lewkowicz and Duguay 1995; Duguay et al. 1997). Also, multi-temporal SAR observations (e.g., ERS-1) have shown to be a particularly effective means for monitoring the seasonal freeze/thaw cycle of boreal forests (Rignot and Way 1994) and subarctic tundra and forest (Duguay et al. 1998). Multidimensional SAR configurations (i.e., multi-frequency, -temporal, -polarization, -incidence angle) are currently being utilized together with optical sensors to improve the approaches developed thus far, so that permafrost maps of other sites in the discontinuous and continuous permafrost zones can be produced. EOS ASTER data used in concert with spaceborne SAR (ERS-1/2 and Radarsat) data will provide the best configuration to date for mapping frozen ground and associated features.

EOS data will enhance the several innovative approaches that have been adopted for sensing permafrost and ground-ice features from space. However, in situ measurements of ground temperatures remain a significant data source for monitoring permafrost conditions and change. For example, long-term programs of monitoring climate-change effects on permafrost and related phenomena in periglacial mountain belts are being carried out in Argentina, Canada, China, Germany, Italy, Japan, Kazakhstan, Norway, Russia, and Switzerland (Haeberli et al. 1995). These and other permafrost data are being entered into a Global Geocryology Database (GGD) to identify, acquire, and disseminate information on permafrost and frozen ground (Barry and Brennan 1993). A pilot GGD is being established at the World Data Center-A (WDC-A) for Glaciology with funding from the U.S. National Science Foundation (NSF). This will enable the archiving of priority Russian data sets and allow the WDC-A to inventory, retrieve, and organize selected priority data sets from other International Permafrost Association members. Increased recognition is also being given to the monitoring of active layer through such initiatives as the Circumpolar Active Layer Monitoring (CALM) project. The available long-term ground-temperature measurements from deep boreholes demonstrate a distinct, but heterogeneous warming trend in lowland permafrost areas (Fitzharris 1996) with some of the largest warmings (2-4˚C/100 yr) occurring over northern Alaska (Lachenbruch and Marshall 1986; Osterkamp 1994).

6.2.3.5 Glaciers and ice sheets

The volumetric mass balance of glaciers or ice sheets is determined by changes to their surface mass balance, basal melting, and, where ice flows into standing water, losses due to calving. The key factors in surface mass balance include mass gains, principally due to winter snow accumulation, and mass losses due to meltwater runoff and sublimation. Small amounts of mass transfer occur through rain freezing on snow, frost accumulation, drifting snow, and avalanches. To the extent that large volumes of ice are involved, change implies global sea-level change, with attendant impacts upon coastal zones. The small ice volumes of individual valley glaciers seem insignificant in this regard until they are grouped together (Meier 1984), but impacts upon local freshwater resources may be highly significant. Also, as noted by Oerlemans (1986), small glaciers can be particularly sensitive to increasing amounts of carbon dioxide in the atmosphere.

The dominant part of change to the volumetric mass balance of a glacier is tied up in its surface mass balance. Data on surface mass balance and volume change are available for about 25 glaciers (Meier 1983, 1984), some of which span more than 50 years since direct measurements first began in 1945. These records are concentrated in the midlatitudes of the Northern Hemisphere, thus rendering them somewhat unrepresentative of global changes to the cryosphere. Midlatitude locations are, however, among those where concentrations of people in search of freshwater resources are likely to be found. Because glaciers are a vital constituent of the freshwater resources of such localities, it is crucial to understand what variations in the surface mass balance imply for the volume of the resource.

There are more-numerous and longer records of glacier advance and retreat (Haeberli 1995), but these are unreliable indicators of volume change. Furthermore, as accurate as good mass-balance measurements in situ can be, the ability to measure volumetric change is moot to water-resource managers if the volume of the ice in question is not known. The significance of change is best understood in relative terms when one is assessing the long-term viability of a resource. There is field evidence that small glaciers in the Rocky Mountains may have lost as much as 75% of the ice volume they had a century ago (Lawby et al. 1995). The volumes of many glaciers, spanning the full size range found in the midlatitudes, must be known in order to assess the significance of such a finding. There is little prospect of accomplishing such a task through field measurement programs.

Much of the current uncertainty about glacier and ice-sheet responses to climate change is generated by the need to make inferences about extensive areas from a few sites where there are field research programs. EOS promises great advances in reducing such uncertainty by virtue of the extensive spatial reach of satellite data. Currently, mass balance can be estimated in the field, either by measuring changes in the thickness and area of grounded ice, or by measuring the seasonal components of the net mass balance. These are often done concurrently with energy exchange and hydrological studies to explain the processes at work. Therefore, existing field research sites, with their potential to calibrate satellite information, are important places in which to concentrate satellite data analysis. Nevertheless, the choice of new sites in which to conduct process studies can benefit from guidance to key areas identified by satellite, and from satellite data on surface characteristics, elevations, ice extent, and atmospheric conditions.

Areal extent and thickness of ice are clearly the key parameters to measure in order to determine volume. The determination of surface area and differentiation into cover classes is a reasonably straightforward task, as illustrated in studies using TM data for Athabasca Glacier (Gratton et al. 1994) and AVHRR data for the Greenland Ice Sheet (Zuo and Oerlemans 1996). The determination of ice-thickness distribution by satellite is a more-challenging problem, though the use of SAR data may eventually prove to be useful here. The possibility that data from ICESat and SAR may prove to be effective for measuring small changes in surface elevation offers prospects for merging satellite data with historical mass-balance records, such as those which have recently been reported for White Glacier (Cogley et al. 1996). Successful merging of the records with satellite data is crucial to the task of using satellites for regional mass-balance assessments of valley glaciers.

Remote sensing has, and will continue to play, an important role in research to improve the understanding and modeling of glacier hydrology. Examples of key contributions are:

1) Accurate representation of topography—DEMs derived from satellite altimetry or SAR interferometry may be crucial as inputs to detailed energy-balance-based melt models. These incorporate effects of slope angle, aspect, and shading on radiation receipts and melt-energy availability. DEMs are also essential for calculation of meltwater routing over and under ice masses.

2) Initialization of melt models—A key issue here is representation of the snow-water-equivalent distribution at the onset of a model run. As yet there is no easy way to find this—on an ice-sheet scale there may be some application for microwave techniques—but current resolution is too coarse at the glacier scale. Surface DEMs may also help define surface-roughness features such as sastrugi, which are likely associated with local variability in snow-water equivalence—and which might determine patchiness of melting snow

cover, which has important implications for albedo and for energy fluxes between adjacent areas of contrasting albedo.

3) Albedo parameterizations—Performance of energy-balance melt models is critically dependent on performance of albedo parameterizations, which have to be applied in a distributed fashion. Remote sensing offers the only realistic prospect for mapping albedo on an ice-mass scale at regular intervals. This is essential for creating databases which will allow development of effective parameterizations—and for testing their performance when ported to sites or years other than those for which they were developed.

4) Meltwater freezing—This is as critical for glacier hydrology as it is for mass balance—especially in the Arctic. Meltwater which refreezes has to be melted at least twice before it runs off—thus adding to effective accumulation. This process delays runoff response to atmospheric temperature changes. Winter warming may not affect melt rates directly, but, by warming surface layers of ice, may reduce the degree of refreezing which occurs, thus promoting more runoff in subsequent summers. Remotely-sensed mapping of distribution of snow/ice facies on ice masses may allow documentation of extent of zones of refreezing—providing a means of testing model performance. Mapping end-of-season snowline positions is also a key means of testing model performance.

5) Glacier outburst hazards—Remote sensing offers a means of locating supraglacial and ice marginal reservoirs, which are potential sources for catastrophic outburst events. It also offers the possibility of recording drainage events, measuring reservoir volume (by comparison of topography of basins in drained and predrainage states), and identifying flood routeways (from surface-collapse features or surface-velocity response) and outburst locations. Long records allow documentation of outburst-event frequency. Large subglacial reservoirs may be detected from high-resolution measurements of surface topography of ice masses, and there may be some prospect for determining major subglacial flow paths the same way.

6) Occurrence of subglacial drainage—Where ice margins are floating, emergence of turbid plumes from ice margins provides evidence of subglacial drainage and helps to locate meltwater efflux points. For land-terminating ice masses, turbidity of meltwater streams may provide some insight into balance of supraglacial/subglacial runoff. In remote areas, remote sensing may allow documentation of distribution of major meltwater input points to oceans—and also provide evidence of timing/duration of these inputs.

6.3 Required measurements, data sets, and parameterizations

6.3.1 Satellite observations

6.3.1.1 Snow cover

A major theme in snow hydrological research over the past decade has been the expanded use of remote sensing for determining snow properties which are used to estimate snow distributions and snowmelt runoff. There has also been a move toward development of physically-based snowmelt models to use with these emerging data, particularly for alpine areas. The coupling of remote sensing and physically-based approaches will enable more-accurate basin-scale forecasts, and will also provide spatially-distributed estimates of snowmelt.

Table 6.2 (pg. 288) summarizes the snow-cover information needed to address the cryospheric-science issues raised in Section 6.2 and to satisfy operational users of EOS data.

Snow extent

Snow-cover-extent information is needed for a wide variety of uses including monitoring and change detection, input to NWP and hydrological models, and for validation of GCMs. The spatial- and temporal-resolution requirements for snow information depend on the application. Continental-scale monitoring of snow cover can be effectively carried out with daily-weekly data at 25-100-km resolution, while spatially-distributed snowmelt and runoff modeling in mountainous terrain requires spatial resolution finer than 1 km. Other useful information can be derived from snow-cover data such as the altitude of the snow line, which can be incorporated into snowmelt forecast models in mountainous areas, e.g., Hartman et al. (1995).

Table 6.2

Parameter Name	Units	Accuracy Need/Avail	Temporal Resolution	Spatial Resolution	Vertical Resolution	Source
Snow extent	—	10%	1/day	30 - 100 m 1/wk, 1/mon	N/A	Radarsat, Landsat
Snow extent	—	10%	1/day, 1 wk, 1/mon	1 km, 12 km, 25 km	N/A	MODIS, AMSR-E, SSM/I
Snow depth	cm	10%	1/day, 1 wk	30 m - 12 km	5 cm	AMSR-E, SSM/I, Radarsat
Snow-water equivalent	mm	10%	1/day, 1 wk	30 - 100 m 12 km	5 mm	Radarsat, AMSR-E, SSM/I
Snow wetness	Yes/ No	—	1/day, 1 wk	30 m - 12 km	—	Radarsat, AMSR-E, SSM/I
Albedo	%	5%	daily	1 km	—	MODIS

Required snow-cover observations to address EOS research and operational requirements. ***(WCRP 1997.)***

NOAA AVHRR data have been routinely used for classification of snow-covered versus snow-free area (Matson et al. 1986; Matson 1991; Xu et al. 1993). Like AVHRR, MODIS will provide near-daily global coverage, but at spatial resolutions ranging from 250 m to 1 km. Only two channels in the visible and near-infrared spectral bands will be available at 250-m resolution; five channels in the visible, near-infrared, and short-wave infrared will be available at 500-m resolution, and the remaining 29 MODIS channels will have a spatial resolution of 1 km, and may not be suitable for snow mapping because they were designed for use over ocean or atmosphere targets. MODIS has onboard visible/near-infrared calibrators while the AVHRR does not, thus we will be able to derive radiances of snow using some of the MODIS sensors. At least one of the visible MODIS channels will not saturate over snow. This will be an advancement over the AVHRR and TM sensors that experience significant saturation over snow and ice targets in the visible channels. The at-launch MODIS snow and ice products will consist of 500-m or 1-km resolution binary maps of snow and ice cover, respectively, produced on a global, daily basis in most months (Hall et al. 1995). In addition, the MODIS cloud mask will be used to derive global maps of cloud cover (Ackerman et al. 1996). After the first EOS launch, techniques initially carried out with Landsat data can be extended to subpixel snow mapping from MODIS and ASTER. But end-member selection needs to be done for each scene. With such analyses, MODIS will also be useful in alpine regions. SAR investigations that are now being pursued with advanced aircraft-mounted systems may be continued using simplified techniques with single-frequency, single-polarization SAR instruments such as ERS-1, the Japanese Earth Remote-Sensing Satellite-1 (JERS-1), and Radarsat to provide useful snow-mapping data.

SSM/I-derived information on snow extent is being used operationally by NESDIS for generation of the weekly snow-cover analysis product. AMSR-E will continue to provide a weekly snow-extent product, which will allow the combination of SMMR, SSM/I, and AMSR-E to provide a continuous time series of snow extent from 1978. The higher spatial resolution of AMSR-E (10 km) relative to SMMR and SSM/I will improve the ability to map snow cover in forested and mountainous regions. AMSR-E snow-cover information will complement snow products from other EOS sensors such as MODIS, Multi-angle Imaging Spectroradiometer (MISR), and ASTER. This will provide EOS researchers with detailed, multi-sensor information on snow-cover extent, albedo, and surface temperature (Chang and Rango 1996).

Snow depth

Snow depth is an important parameter for input to NWP snow-cover analyses and for validating GCM simulations. For example, the most frequently used GCM snow-validation standard is the United States Air Force (USAF) global snow-depth data set of Foster and Davy (1988). This data set is based on in situ snow-depth observations, which are not available in many sparsely-populated areas of the world. A global snow-depth map will be produced as a special product from AMSR-E (Chang and Rango 1996). This will most likely apply a density climatology to the AMSR-E snow-water-equivalence map product since snow density exhibits considerably less spatial and temporal variability than snow-water equivalence. The accuracy of this product will depend on the success of snow-water-equivalence-algorithm development activities to account for forest and land-cover effects on brightness temperatures.

Snow-water equivalence

Regular observations of snow-water equivalence are needed to satisfy a wide range of requirements in research and operations, e.g., validation of climate and hydrological models (e.g., AMIP), water resource monitoring (e.g., Goodison and Walker 1994), and input to flood-forecast models. Accurate basin-wide snow-water-equivalence estimates are especially critical in the western U.S. where much of the annual runoff and groundwater recharge comes from melting of the mountain snow pack each spring. The lack of a reliable global snow-water-equivalence climatology for validating GCM climate simulations is a critical data gap that also needs to be addressed. Determining snow-water equivalence directly from remotely-sensed data has been an objective within the remote-sensing community for some time; although operational gamma radiation methods have been developed for areas with gentle topography, aircraft flight height constraints and gamma signal extinction problems in smaller, rugged mountain basins with deep snowpacks generally prevent the use of this method in these areas. SAR at multiple frequencies and polarizations has yielded promising results for snow-water-equivalence measurement (Shi et al. 1990), but operational satellites do not provide the necessary data. At present, the measurement of the spatial distribution of snow-water equivalence and total snow volume within a basin must be performed by intensive field sampling to attempt to represent the large spatial variability of mountain snowpacks. Logistical and safety limitations generally restrict the number of field samples that may be so obtained (Elder et al. 1991). Thus, the problem of determining the volume and distribution of snowpack water storage within mountain basins remains acute.

Estimation of distributed snow-water equivalence is challenging because of the many factors that affect its distribution, and the small correlation length of the snow-water-equivalence spatial distribution. Further, difficulties associated with accurately determining the time of maximum accumulation present a problem for snowmelt-runoff forecasters. The simplicity of regression models makes them an attractive means of estimating snow-water equivalence because of the large amount of work required to make direct measurements of snow-water equivalence on the catchment scale. An approach to modeling spatially-distributed snowmelt in steep, alpine basins was proposed using net potential radiation, distributed across the basin using a DEM, as the main factor determining relative snowmelt (Elder et al. 1991). Such an approach enables the use of a detailed, physically-based snowmelt model for each physically different subregion of the basin at the scale of interest. Testing this approach on the 1.2 km^2 Emerald Lake basin in California's Sierra Nevada suggests that little information is lost in going from a 5-m to 25-m grid, but that use of a 100-m grid may result in significant inaccuracies (Bales et al. 1992).

Remote sensing allows estimation of several of the important hydrologic variables for snowmelt modeling from space or aircraft. From Landsat one can map snow at subpixel resolution, to an accuracy as good as with aerial photographs. From hyperspectral sensors (currently the Airborne Visible and Infrared Imaging Spectrometer [AVIRIS]) one can estimate snow-grain size, albedo, liquid-water content in the surface layer, and subpixel coverage. Using two-frequency, co-polarized SAR, one can map snow through thick cloud cover to an accuracy of 80%, and estimate liquid-water content to about 2%. Work on estimation of snow-water equivalence is continuing, with promising results from the Shuttle Imaging Radar-C (SIR-C)/X-Band Synthetic Aperture Radar (X-SAR), from photogrammetry, and from snowmelt modeling with time-series SCA data. A fully automated method of subpixel snow-cover mapping uses Landsat TM data to map snow cover in the Sierra Nevada and make quantitative estimates of the fractional snow-covered area within each pixel (Rosenthal and Dozier 1996). TM scenes are modeled as linear mixtures of image end-member spectra to produce the response variables for tree-based regression and classification models. The algorithm has been tested on a different TM scene and verified with high-resolution, large-format, color aerial photography. Snow-fraction estimates from the satellite data can be as accurate as those attainable with high-resolution aerial photography, but they are obtained faster, at much lower cost, and over a vastly larger area.

The development and validation of snow-water-equivalence algorithms for use with AMSR-E is a major EOS activity. A detailed discussion of the snow-water-equivalence-algorithm development and validation process for AMSR-E is provided in Chang and Rango (1996). Figure 6.4 documents the multiple steps required to generate a snow-water-equivalence estimate. Detailed field observations of snow properties and corresponding microwave signatures over a variety of terrain and land-cover categories are an essential component of this process. This is an important contribution of snow research activities within the CRYSYS IDS (Goodison and Brown 1997).

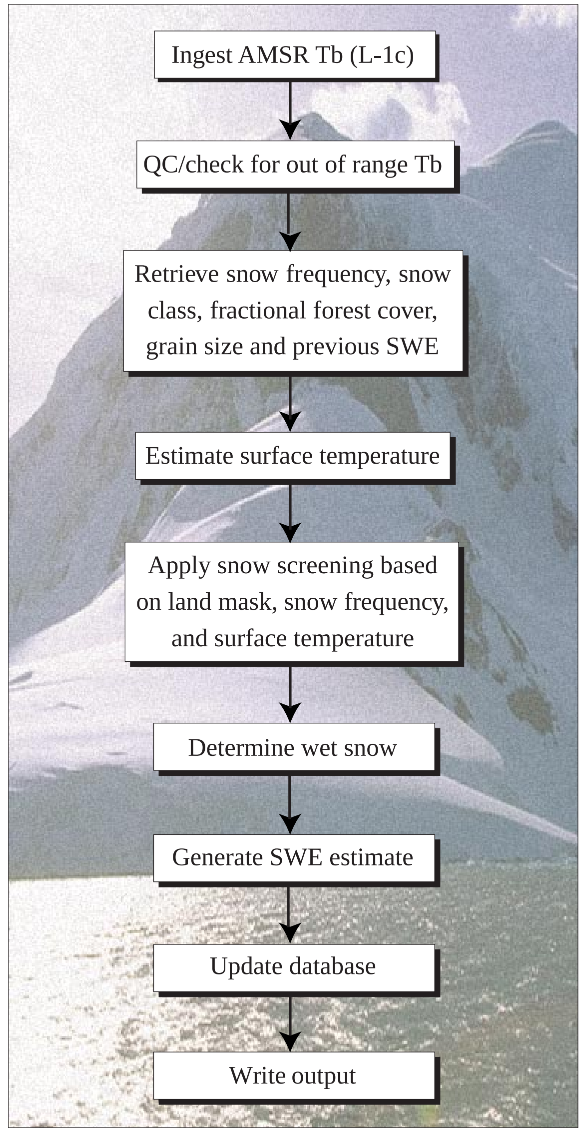
It is also possible to infer snow-water equivalence after the fact from measurements of snow-cover depletion. With a time series of snow cover, e.g., from TM imagery, one can tell when the snow cover disappears, i.e., when snow-water equivalence goes to zero. Then, using a spatially-distributed snowmelt model, one can back calculate from the time snow cover disappears at a point, and then infer the starting value of snow-water equivalence. This method has been implemented using TM scenes for a small watershed in the Sierra Nevada, California (Cline 1997). In this first test, the inferred initial snow-water equivalence agreed well with snow-survey measurements. The reverse approach can be used with a dense time series of remote-sensing scenes, such as will be available from MODIS, to forecast potential snowmelt; such a forecasting approach could set aside the need to estimate snow-water equivalence, and would rely on properties more readily measured from remote-sensing platforms, including SCA and reflectivity.

Snow wetness and albedo

Snow wetness provides important hydrological information such as the onset of melt, and may be a particularly sensitive indicator of change given that climate warming would likely be associated with more frequent rain-on-snow events and more frequent thaw events. Snow-wetness information can be derived from both active and passive microwave data.

Snow-surface albedo is needed for a multitude of uses including climate-change monitoring (e.g., Robinson et al. 1993), understanding snow-cover-climate feedbacks (e.g., Groisman et al. 1994a), input to NWP and hydrological models, and validation of GCMs. Global coverage of snow albedo at 1-km resolution will be provided by MODIS. For finer resolution requirements such as snow ablation in hydrological models, surface-albedo information can be derived from ASTER and the Enhanced Thematic Mapper+ (ETM+) for defined target areas at a horizontal resolution of 15-30 m over the visible and infrared portions of the spectrum.

Figure 6.4



Schematic flow diagram of AMSR-E SWE algorithm. ***(Chang and Rango 1996).***

6.3.1.2 Sea ice

Sea-ice models, especially those used in climate simulations, treat many processes in a very idealized manner. The reasons include inadequate knowledge of the physics involved, inadequate observational data to define certain key parameters, and the computational expense of including more-realistic treatments. Some process parameterizations which are particularly in need of improvement are listed below, along with the associated observational requirements.

Sea-ice momentum balance—As indicated in Section 6.1.3, GCMs often ignore sea-ice motion, although the trend is now to include a sea-ice dynamical scheme. Physical parameters that arise in such models are shear and compressive strength and drag coefficients. Since strength parameters cannot be measured directly, they are generally inferred by comparisons of observed and modeled buoy drift (e.g., Hibler and Walsh 1982; Flato and Hibler 1992, 1995). More-detailed ice-motion fields, such as those provided by sequential satellite imagery, should allow discrimination between various proposed strength parameterizations, more specifically, the shape of the plastic-yield curve (see, e.g., Ip et al. 1991; Ip 1993). Observations of ice roughness, which is dominated by ridging intensity, may allow more-realistic drag parameterization, which in turn affects a model's ability to reproduce observed ice drift. Required observations include ice motion and deformation. The accuracy and resolution requirements, along with the data sources, are provided in Table 6.3. Some encouraging results on observed ice drift have recently been obtained through wavelet analysis of satellite ERS-1 SAR images (Liu et al. 1997), and the wavelet technique is now being applied by Liu to the passive microwave data of the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave Imager (SSM/I).

SEB—Because of its important role in the SEB and climate feedback, it is crucial to parameterize sea-ice albedo accurately. Such parameterization is confounded by both spatial inhomogeneity and the dramatic drop in albedo accompanying snow melt and surface-melt-pond formation in spring (e.g., Morassutti 1989; Barry 1996). More-complete parameterizations, which explicitly include melt ponding, are being developed (Ebert and Curry 1993), and large-scale estimates of surface albedo based on DMSP and AVHRR satellite observations (Scharfen et al. 1987; Lindsay and Rothrock 1994) should aid in this development. Other terms in the sea-ice surface-energy budget are also uncertain, and the available parameterizations are in need of more-comprehensive validation. Data sets required to develop and validate parameterizations of the surface-energy-exchange process include profiles of atmospheric quantities (for radiation calculations), surface quantities like albedo, temperature, and roughness, and ice concentration. Specific requirements are provided in Table 6.4.

Ice mass balance—Continuous deformation, growth, and melt ensure that a region of sea ice contains a range of ice thickness from open water through newly-formed and multi-year ice, to pressure ridges tens of meters thick. Although a theoretical framework to describe this thickness distribution has been available for some time (Thorndike et al. 1975), it has not been widely used owing to computational complexity and uncertainty regarding key parameters. On the other hand, ocean-atmosphere heat exchange is very sensitive to details of the thin portion of the thickness distribution (Maykut 1982), and the time scale of temporal variability in Arctic ice volume is dominated by the long survival time of thick ridged ice (Flato 1995). Hence, some account of the thickness distribution should be made in climate models. The most common approach so far is to consider only two thickness categories, open water and the mean thickness (e.g., Hibler 1979; Parkinson and Washington 1979; Flato and Hibler 1992), but this is rather crude. Parameterizations required to describe ridge redis-

TABLE 6.3

PARAMETER NAME	UNITS	ACCURACY NEED/AVAIL	TEMPORAL RESOLUTION	SPATIAL RESOLUTION	VERTICAL RESOLUTION	SOURCE
Ice displacement	m	1 km/300 m	2/week	5 km	N/A	MODIS, Radarsat, buoys
Ice deformation	s^{-1}	.5% / .1%	2/week	5 km	N/A	MODIS, Radarsat, buoys

Accuracy and resolution requirements to observe sea-ice momentum balance *(WCRP 1997).*

Table 6.4

Parameter Name	Units	Accuracy Need/Avail or Absolute::Rel	Temporal Resolution	Spatial Resolution or Horizontal Resolution::Cover	Vertical Resolution	Source
Sea-ice concentration	%	7%	daily	20 km	—	AMSR-E
Floe-size statistics	—	10%	weekly	100 km	—	MODIS, Radarsat
Ridge statistics	—	10%	weekly	100 km	—	Radarsat
Cloud fraction	—	4% / 30%	daily	1 km	—	MODIS, climatology, stations
Cloud optical depth	—	15%	daily	1 km	—	MODIS, aircraft
Cloud particle phase	—	—	daily	1 km	—	MODIS, aircraft
Cloud effective particle radius	—	25%	daily	1 km	—	MODIS, aircraft
Cloud top temperature	—	6%	daily	1 km, 100 km	—	AIRS, AMSU, MODIS
Cloud top pressure	—	6%	daily	100 km	—	AIRS, AMSU, HSB
Atmos. ice crystal precip.	—	—	—	—	—	ICESat, aircraft
Atmospheric aerosol	—	—	—	—	—	ICESat, aircraft
Surface albedo	—	0.05 / 0.1	daily	1km	—	MODIS, climatology
Ice surface	—	1 K / 2 K	daily	1 km, 100 km	—	MODIS, AIRS, buoys, stations
Surface roughness	—	—	—	—	—	laser, survey
Humidity profile	g/kg	10% (goal)::5%	2/day (d, n)	50 x 50 km::G	2 km::Atmos	AIRS (05), HSB
Temperature profile	K	1.0 K::0.4 K	2/day (d, n)	50 x 50 km::G	2 km::Atmos	AIRS (07)

Data sets required to develop and validate parameterizations of the surface-energy-exchange process *(WCRP 1997).*

tribution and strength in the Thorndike et al. (1975) scheme have been investigated numerically by Flato and Hibler (1995). Most of the observations used in such studies are from submarine upward-looking sonar profiles. However, some applicable remote-sensing observations are being assembled from sequential SAR imagery (Kwok et al. 1995; Stern et al. 1995). The quantities of interest here are the ice-thickness distribution, concentration and type, snow thickness, and ridge and floe statistics. Specific requirements are provided in Table 6.5.

Observing the fraction of open water within the ice pack is a priority, both in terms of monitoring and understanding variations which might occur, and in terms of improving the representation of open-water effects in sea-ice and climate models. Since open-water fraction is simply one minus the ice concentration, observations from passive microwave satellites are in principle available from the mid 1970s onward. However, the errors inherent in concentration estimates, roughly 2-7% (Cavalieri 1992), imply errors in open water estimates of nearly 100%! Improving these errors will be a challenge for scientists using future EOS sensors. Nevertheless, the long time series of existing ice-concentration data (Zwally et al. 1983; Parkinson et al. 1987; Gloersen et al. 1992) provide a useful measure of seasonal and interannual variability in open-water fraction.

6.3.1.3 Lake ice

The freeze-up/break-up of lakes can be readily monitored by satellite. Barry and Maslanik (1993) describe the potential use of passive microwave data for monitoring large water bodies. Small lakes can be readily mapped from visible and infrared data for cloud-free conditions, which are generally more common in spring. Dates determined by remote sensing generally compare well with surface observations (Wynne et al. 1996). SAR data have also been used to study lake-ice processes (freeze-up/break-up, freezing of ice to the lake bed, and ice thickness [Hall et al. 1994; Jeffries et al. 1994; Morris et al. 1995; Jeffries et al. 1996; Duguay and Lafleur 1997]).

With EOS instruments we can exploit the synergism between optical and microwave spaceborne imagery for lake-ice monitoring. As shown in Table 6.6, satellite data should ideally be acquired on a daily basis to determine, with as much precision as possible, the timing of fall freeze-up and spring break-up. Although the sensors identified in this table are well suited for lake-ice studies on large lakes, they lack the spatial resolution needed to

Table 6.5

Parameter Name	Units	Accuracy Need/Avail or Abs::Rel	Temporal Resolution	Horizontal Resolution ::Cover	Vertical Resolution ::Cover	Comments
Ice thickness distribution	m^{-1}	thickness: 10% / 50%	—	—	—	sonar, models, Radarsat
Snow depth	m	5 cm / 20 cm	—	—	—	climatology
Sea ice	%	7%::	—	12 km :: Ocean/Cryo	N/A :: Sfc	AMSR-E (11)
Sea-ice type	%	11%::	—	12 km :: Ocean/Cryo	:: Sfc	AMSR-E (12)

Data sets and accuracy required for estimating sea-ice mass balance *(WCRP 1997).*

Table 6.6

Parameter Name	Units	Accuracy	Temporal Resolution	Spatial Resolution	Vertical Resolution	Source
Freeze-up/ break-up date	Julian Day	1 day	daily	1 km 5 km		MODIS, AMSR-E, SSM/I

Required lake-ice observations for cryospheric monitoring and change detection *(WCRP 1997).*

Table 6.7

Parameter	EOS Instrument	Accuracy Absolute:: Relative	Temporal Resolution	Horizontal Resolution[2]	Vertical Resolution	Comments
Land cover / land use	ASTER, ETM+	10%::	1/year (late July - early August)	15 - 30 m (VIS, NIR, IR)	—	SPOT (useful), MODIS and AVHRR (resol. limiting)
Snow cover	ASTER, ETM+	10%::	8/year (mid-March to mid-June)	15 - 30 m (VIS, NIR, MIR)	—	ERS, JERS, Radarsat, SPOT (useful). AMSR, MODIS, and SSM/I (resol. limiting)
Snow depth	ASTER, ETM+	—	8/year (mid-March to mid-June)	15 - 30 (VIS, NIR, MIR)	—	ERS, JERS, Radarsat (useful). AMSR, SSM/I (resol. limiting)
Soil moisture	ASTER, ETM+	—	6/year (mid-June to late August)	15 - 90 m (VIS, NIR, MIR, TIR)	—	ERS, JERS, Radarsat (useful). AMSR, MODIS, AVHRR (resol. limiting)
Surface reflectance and albedo	ASTER, ETM+	4%::1%	8/year (early May to late August)	15 - 30 m (VIS, NIR, MIR)	—	MISR, MODIS, AVHRR (resol. limiting)
Surface brightness temperature	ASTER, ETM+	1-2 K :: 0.3 K	8/year (early May to late August)	60 - 90 m (TIR)	—	AMSR, MODIS, AVHRR, SSM/I (resol. limiting)
Surface kinetic temperature	ASTER	1-4 K :: 1%	8/year (early May to late August)	90 m (TIR)	—	MODIS (resol. limiting)
Digital elevation model	ASTER	5-10 m ::	1/year (mid-July to early August)	15 m (PAN)	1 m	SPOT and SAR (useful)
Permafrost surface displacement features	ASTER, ETM+	—	1/year (mid-July to early August)	15 - 30 m (VIS, NIR, MIR)	—	SPOT, ERS, JERS, and Radarsat (useful)
Freeze / thaw cycle (vegetation/soil)	—	—	1/week (complete year)	—	—	ERS, JERS, Radarsat, (useful) AMSR and SSM/I (resol. limiting)

Note: Permafrost extent and active-layer thickness cannot currently be extracted directly using remotely-sensed data. (Instruments offering spatial resolutions coarser than 100 m may not be useful for most permafrost investigations.)

Required satellite observations for monitoring and change detection of permafrost[1] and seasonally-frozen ground *(WCRP 1997).*

monitor freeze-up and break-up on small (shallow) lakes, which are found at many locations in subarctic and arctic regions. Radarsat and ERS-2 SAR sensors, and ASTER visible/infrared spectral bands are seen as important data sources for lake-ice-process studies on shallow lakes, even though the temporal resolution is of the order of one week to a month (depending on the latitude of the study site and number of incidence angles available). However, as indicated by Morris et al. (1995), shallow lakes represent promising sites for long-term monitoring and the detection of changes related to global warming and its effects on polar regions.

6.3.1.4 Frozen ground and permafrost

As shown in Table 6.7, monitoring of permafrost conditions requires the observation of several parameters. Some of these (local environmental factors) are required as input or transfer functions in ground thermal models. Because permafrost conditions are strongly controlled by local factors, and permafrost features (i.e., surficial expressions) are also local in scale, ASTER will provide the best overall configuration for permafrost mapping and monitoring efforts (i.e., high-spatial resolution, stereo capabilities, and wide range of spectral bands), followed by ETM+ to be flown on Landsat-7.

Local environmental factors—Local factors (i.e., vegetation cover, snow cover, elevation/slope/aspect, and soil-moisture conditions) commonly override the influence of larger scale macroclimatic factors on ground thermal conditions. These factors will be derived from ASTER to complete ongoing regional permafrost mapping efforts at several permafrost study sites (i.e., mapping the presence/absence and active-layer depth), and to determine their influence on the thermal regime of permafrost (by coupling the ASTER-derived variables to a heat transfer model). While Visible and Near Infrared (VNIR) (including the stereo capabilities for the production of DEMs) and short-wavelength infrared (SWIR)

TABLE 6.8

PARAMETER NAME	UNITS	ACCURACY NEED/AVAILABLE	TEMPORAL RESOLUTION	SPATIAL RESOLUTION	VERTICAL RESOLUTION	SOURCE
Ice surface topography	m	0.5 - 5 m /?	yearly	10 - 100 m	1 - 10 m	ICESat, SAR
Area covered by ice	km^2	0.01 - 0.5 km^{-2} /0.001 km^{-2}	yearly	100 m - 1 km	—	ASTER, AVHRR, ETM, MODIS, TM
Ice-thickness distribution	m	0.5 - 5 m /?	yearly	100 m - 1 km	1 - 10 m	SAR
Ice-flow velocity	m d^{-1}	0.05 - 0.5 m d^{-1} /?	daily, yearly	0.1 - 1 m	—	ICESat, SAR
Seasonal ice/ snow cover	km^2	0.01 - 0.5 km^{-2} /0.001 km^{-2}	daily, yearly	10 m - 1 km	—	ASTER, AVHRR, ETM, MODIS, TM
Snow-depth distribution	m	- 0.25 m	daily, yearly	10 m - 1 km	0.05 - 0.5 m	ASTER, SSM/I
Snow-water equivalent	mm	2.5 - 25 mm	daily	10 m	5 - 50 mm	SAR
Surface temperature	K	0.5 K / 2 K	daily	10 m - 1 km	—	ASTER, AVHRR, ETM, MODIS, TM
Albedo distribution Cloud cover	% %	1% /5-10% 10%	daily daily	10 m - 1 km 1 - 10 km	— —	ASTER, AVHRR, ETM, MODIS, TM, CERES
Meltwater runoff	Yes /No	— /—	daily	1 - 100 m	—	ASTER, TM, ETM

Data sets and accuracy required to monitor glacier and ice-sheet responses to changing climate *(WCRP 1997).*

channels are planned to be used to derive the local factors, the thermal infrared (TIR) channels will be utilized to test the heat transfer model. The ground-surface temperature in such models is usually approximated using N-factors (factors applied as transfer functions between the air temperature and the temperature at the ground surface, to account for the local influence of vegetation and snow cover). The TIR channels will likely provide an improvement over the current N-factors approximation. SAR sensors, such as those onboard Radarsat, will also be useful, in particular, for deriving information on snow and soil-moisture conditions.

Permafrost features—Features indicative of permafrost degradation such as thaw lakes, active-layer detachment slides, and retrogressive thaw slumps will be mapped and monitored (VNIR and SWIR channels in particular). These features are particularly good indicators of climate warming. Other surface features, known to have an impact on permafrost conditions, such as new human settlements and forest fires (the after effect, i.e., burned areas), can also be mapped with ASTER. Active microwave SAR imagery acquired by other non-EOS sensors is also likely to be useful (whether used alone or in combination with ASTER data) for mapping thermokarst features.

Seasonally frozen ground—Global warming may significantly alter the duration of the freeze-thaw periods at high- and mid-latitude locations. A modification of the length of the frost-free period could have significant ecological, hydrological, engineering, and economical implications. Active and passive microwave imagery have been shown to be particularly useful for monitoring the freeze-thaw cycle of the landscape (Table 6.7).

6.3.1.5 Glaciers and ice sheets

Surface topography, area of ice cover, the spatial distribution of its thickness, surface flow velocity, areas covered by snow and exposed ice during melt periods, snow-depth distribution, its water equivalent, the distributions of surface temperature and albedo, and the presence or absence of meltwater runoff are all important parameters to use in identifying the responses of glaciers and ice sheets to climatic change. They are also needed to assess implications of these responses for sea-level rise and regional water resources. The temporal, spatial, and vertical resolutions stated in Table 6.8 vary according to the scale of the ice involved and the task at hand. The lower ends of the spatial- and vertical-resolution ranges apply to small valley glaciers; the upper ends to large glaciers and ice sheets. Yearly resolution is needed for monitoring responses to climatic change. Daily resolution is ideal to estimate mass-balance responses to short-term weather variations, such as winter accumulation in the snow pack and summer release of meltwater runoff to streams. Daily resolution, however, with acceptable spatial resolution for small glaciers, cannot be obtained with current satellite scanning systems.

The EOS experiment most closely associated with ice-sheet mass balance is ICESat. By accurately measuring ice surface topography, improved ice-flow models are possible. By measuring it repeatedly, changes in surface elevation are revealed. After correction for small changes caused by evolution of snow-density profiles and of bedrock motions, the residual is the change in ice-sheet mass balance and hence the contribution of ice sheets to sea-level change. Such experiments are aided by radar altimetry, but laser precision is needed to allow accurate, ice-sheet-wide assessment of mass balance over years rather than decades. Repeat laser altimetry will also identify those regions that are changing especially rapidly, thus identifying important locations for process studies.

Landsat TM data, available in six shortwave bands with spatial resolution of 30 m and one thermal band with 120-m resolution since 1972, are an important source of glacier information pertaining to area covered by ice, as well as its differentiation into seasonal ice/snow cover, so they play a prominent role in Table 6.8. AVHRR has an advantage over TM in that it can provide daily resolution, but its spatial resolution of 1.1 km makes it most suitable for use over large ice sheets (Zuo and Oerlemans 1996). MODIS, with near-daily temporal resolution over a few bands at 250-to-500-m spatial coverage, may prove useful in modeling meltwater discharge from large valley glaciers and small ice sheets, thus providing an effective blend of spatial and temporal resolution.

Increased spatial and spectral resolution over that of the TM is to be gained from the ETM and from ASTER. ASTER, with 14 bands in the visible, near-infrared, and TIR, will have the added benefit of monochromatic stereo imaging. Furthermore, the estimation of snow-depth distribution is possible from ASTER, as it is with SSM/I. It should also be possible to determine the seasonal timing of meltwater runoff from most ice-covered areas with the enhanced resolution.

The spectral characteristics of the foregoing sensors provide the keys to detailed mapping of the components of radiative exchange on glaciers (Gratton et al. 1994) because their data are useful for determining surface temperature and the albedo distribution. Furthermore, it is possible to determine separate albedo values for the visible and near-infrared parts of the solar spectrum, and to validate them against similar measurements in situ over ice (Cutler and Munro 1996) and snow (Cutler and Munro 1996; Marks and Dozier 1992). As indicated in earlier modeling work by Munro and Young (1982), and shown more recently in situ by Van de Wal et al. (1992), albedo is a powerful determinant of the ablation which supplies meltwater at the ice surface. If such information is to be used effectively in regional hydrological modeling, then it will also be necessary to have satellite cloud-cover data as well, such as can be provided by the Clouds and the Earth's Radiant Energy System (CERES).

Since the 1991 launch of the ERS-1 satellite and its C-band SAR, glacier monitoring has been possible to spatial resolutions of approximately 30 m through cloud cover and darkness. Since October 1995, Radarsat has offered SAR data with 9-m resolution. SAR data reveal information about subsurface glacier processes and features, imaging buried crevasses and the location of the equilibrium line through winter snow cover. SAR's capability of looking so deeply into the ice suggests possibilities for measuring ice-thickness distribution as well. Another unique attribute of SAR is that careful repeat imaging spaced a few days or weeks apart can be used in an interferometric mode. Under optimum conditions, radar interferometry allows precise ice-flow-velocity determinations over short intervals. There is also the possibility that SAR will one day yield the snow-water equivalent of the snow pack as it changes with the seasons, but satellites do not yet provide the necessary data.

6.3.2 Related international science programs

6.3.2.1 The Boreal Ecosystem-Atmosphere Study (BOREAS)

is a large-scale international interdisciplinary experiment in the northern boreal forests of Canada. Its goal is to improve our understanding of the boreal forests—how they interact with the atmosphere, how much CO_2 they can store, and how climate change will affect them. BOREAS wants to learn to use satellite data to monitor the forests and to improve computer simulation and weather models so scientists can anticipate the effects of global change (http://boreas.gsfc.nasa.gov/).

6.3.2.2 ACSYS

ACSYS is a WCRP project aimed at improved understanding of the role of the Arctic in the climate system, and of the effects of global climate change and variability on the Arctic. Specific scientific goals of ACSYS include improving the representation of the Arctic in climate models, developing plans for monitoring Arctic climate, and determining the role of the Arctic in climate sensitivity and variability. These scientific goals are closely aligned with EOS goals. Of particular interest to EOS are in situ observations of Arctic Ocean hydrography, cloud-radiation interactions, observations of ice thickness and Lagrangian motion, and the assembly of historical oceanographic and hydrological data sets (many of which are currently held in Russia). More information can be obtained from WCRP (1994) or via the ACSYS web page (http://www.npolar.no/acsys/).

6.3.2.3 Surface Heat Budget of the Arctic Ocean (SHEBA)

SHEBA is a multi-national project whose goals are to develop and test models of Arctic ocean-ice-atmosphere interactions and to improve the interpretation of satellite remote-sensing data in the Arctic. Both of these will be based in large part on detailed observations to be made during the 1997-1998 field experiment from the ship frozen in the Beaufort Sea and from associated camps. The very comprehensive annual-cycle data set to be collected during the field program will be especially valuable to EOS for calibration and validation of remote-sensing algorithms and for improving model parameterizations of various energy-exchange processes. More details are available in Moritz et al. (1993) or via the SHEBA web page (http://sheba.apl.washington.edu/).

6.3.2.4 International Arctic Buoy Program (IABP)

IABP oversees and monitors a network of automatic data buoys which have been repeatedly deployed across the Arctic since 1979. These buoys measure surface pressure, temperature, and ice motion. Climatological and synoptic data from the IABP buoys provide information on large-scale winds, ice drift, and deformation. Further details can be found via the IABP web page (http://iabp.apl.washington.edu/).

6.3.2.5 Submarine Arctic Science Cruise (SCICEX) program

The SCICEX program is a five-year multi-agency program that makes use of U.S. Navy nuclear submarines as research platforms. The aim of this program is to increase our fundamental understanding of processes in the Arctic Ocean. The Office of Naval Research (ONR) and the NSF co-fund this unclassified basic research mission. Further details can be found via the web page (http://www.nsf.gov/od/opp/arctic/logistic/whatsnew.htm#scicex).

6.3.2.6 Program for Arctic Regional Climate Assessment (PARCA)

PARCA is a NASA project formally initiated in 1995 by combining into one coordinated program various investigations associated with efforts starting in 1991 to assess whether airborne laser altimetry could be applied to measure ice-sheet-thickness changes. It has the prime goal of measuring and understanding the mass balance of the Greenland ice sheet. The main components of the program are:

- Periodic, airborne-laser-altimetry surveys along precise repeat tracks across all major ice-drainage basins. The first survey was completed in 1993/1994, with repeat flights along selected routes in 1995 and 1996, when flights were also made over ice caps in eastern Canada, Svalbard, and Iceland.

- Ice-thickness measurements along the same flight lines.

- Monitoring of various surface characteristics of the ice sheet using satellite radar altimetry, SAR, passive microwave, AVHRR, and scatterometer data.

- Surface-based measurements of ice motion at 30-km intervals along the 2000-meter contour completely around the ice sheet, with interpolation of local relative ice motion using interferometric SAR.

- Shallow ice coring (10-200 meters) at many locations to infer recent climate history, atmospheric chemistry, and interannual variability of snow-accumulation rates and to measure temperature and vertical ice motion at various depths.

- Investigations of SEB and factors affecting snow accumulation and surface ablation. This program is a collaborative effort with NSF and includes the installation of automatic weather stations (AWS) at the deeper drill-hole sites.

- Estimating snow-accumulation rates by model analysis of column water vapor obtained from radiosondes and TIROS Operational Vertical Sounder (TOVS) data.

- Detailed investigations of individual glaciers and ice streams responsible for much of the outflow from the ice sheet.

- Development of a thermal probe to measure various ice characteristics at selected depths in the ice sheet.

- Continuous monitoring of crustal motion using Global Positioning System receivers at coastal stations.

6.4 EOS contributions

The cryosphere plays a significant role in the global climate and hydrologic systems over a wide range of time and space scales, and cryospheric feedback processes dominate the high-latitude response of GCM climate-change experiments. Furthermore, snow cover and mountain glaciers provide important sources of freshwater and hydro-electric power for large populations, particularly in the western United States, Canada, northern Europe, and parts of Asia. Changes in the mass balance of the major ice sheets also have the potential for significant impacts on global sea level over the next few centuries.

The cryosphere, through its interactions with other components of the Earth system, contributes to the feedback processes that modify the global and regional response to climate change. At the same time, global climate change is likely to produce cryospheric changes that will have significant ecosystem and societal impacts at the regional scale. Understanding and quantifying cryosphere-climate interactions, improving their representation in numerical models, and monitoring changes in cryospheric parameters are, therefore, the objectives of several international research programs. EOS will provide a major contribution to all three areas.

Several EOS instruments (e.g., AMSR-E, ASTER, GLAS on ICESat, MODIS) will enable improved monitoring and measurement of the cryosphere from space. These instruments either improve on current measurement capabilities or, as in the case of GLAS, will provide measurements that are not possible on a routine basis with current systems.

Examples of the type of research that will be possible with EOS are seen in the EOS IDS investigations that have a particular cryospheric focus:

- Use of CRYSYS to Monitor Global Change in Canada. Principal Investigator, Barry E. Goodison (http://www.tor.ec.gc.ca/CRYSYS/).

- Polar Exchange at the Sea Surface (POLES): The Interaction of Ocean, Ice, and Atmosphere. Principal Investigator, D. Andrew Rothrock (http://psc.apl.washington.edu/POLES/).

- Hydrology, Hydrochemical Modeling, and Remote Sensing in Seasonally Snow-covered Alpine Drainage Basins. Principal Investigator, Jeff Dozier (http://www.icess.ucsb.edu/hydro/hydro.html).

- Global Water Cycle: Extension across the Earth Sciences. Principal Investigator, Eric J. Barron (http://eoswww.essc.psu.edu/gwchome.html).

- Interdisciplinary Determination of Snow Accumulation Patterns on the Greenland Ice Sheet: Combined Atmospheric Modeling and Field and Remote-Sensing Studies. Principal Investigator, Robert Bindschadler.

- In addition, the National Snow and Ice Data Center (NSIDC), is one of eight archives participating in the Earth Observing System Data and Information System (EOSDIS) as a Distributed Active Archive Center (DAAC). The NSIDC DAAC exists to help broaden understanding of snow and ice, of their properties, characteristics, and contexts, and of their significance for human activity. The Center archives analog and digital snow and ice data, creates and distributes data products, and maintains a large library collection in support of snow and ice research. Established in 1982, NSIDC is co-located with the WDC-A for Glaciology and, through the Polar DAAC Advisor Group (PoDag), serves an important coordinating role in snow and ice research (http://www-nsidc.colorado.edu/NASA/GUIDE/index.html).

The enhanced measurement capability represented by EOS will improve our ability to monitor and forecast changes in cryospheric parameters that have significant societal consequences. EOS products such as snow-water equivalence are important for water-resource planning. Similarly, information on sea-ice concentration is significant for ship routing and fisheries. EOS data will realize other major benefits to society by helping to reduce the uncertainties associated with climate change. EOS data and EOS-sponsored research activities will improve the representation of cryospheric processes in hydrological and climate models (e.g., mountain snowpacks and sea-ice dynamics), and improve our understanding of the coupling between atmosphere-ocean-cryosphere and the combined impact of cryospheric changes on sea level. Cryospheric data sets currently available for model boundary conditions or validation will be greatly enhanced by EOS satellite data and EOS-related products (Barry 1997a, b).

References

Abdalati, W., and K. Steffen, 1995: Passive microwave-derived snow melt regions on the Greenland ice sheet. *Geophys. Res. Lett.*, **22**, 787-790.

Ackerman, S., K. Strabala, P. Menzel, R. Frey, C. Moeller, L. Gumley, B. Baum, C. Schaaf, G. Riggs, and R. Welch, 1996: *Discriminating Clear-Sky from Cloud with MODIS*. Algorithm Theoretical Basis Document ATBD-MOD-06, NASA Goddard Space Flight Center, 119 pp. [*http://modarch.gsfc.nasa.gov/MODIS/ATBD/atbd_mod06.pdf*].

Alley, R. B., and I. M. Whillans, 1984: Response of the East Antarctic ice sheet to sea-level rise. *J. Geophys. Res.*, **89**(C), 6487-6493.

Alley, R. B., 1989: Water pressure coupling of sliding and bed deformation: I. Water system. *J. Glaciology*, **35**, 108-118.

Alley, R. B., and I. M. Whillans, 1991: Changes in the West Antarctic ice sheet. *Science*, **254**, 959-963.

Alley, R. B., 1992: Flow-law hypotheses for ice-sheet modeling. *J. Glaciology*, **38**, 245-256.

Alley, R. B., 1996: Towards a hydrologic model for computerized ice sheet simulations. *Hydrological processes*, **10**, 649-660.

Alley, R. B., 1997: Antarctica and sea-level change. Antarctic Journal of the U.S. (in press).

Anderson, E. A., 1976: *A point energy balance model of a snow cover*. Office of Hydrology, National Weather Service, NOAA Tech. Rep. NWS 19, 150 pp.

Armstrong, R. L., and M. Hardman, 1991: Monitoring global snow cover. *Proc. Western Snow Conference*, 59th Annual Meeting, 103-108.

Armstrong, R. L., and A. Krenke, 1997: *Former Soviet Union Hydrological Snow Surveys, 1966-1990*. National Snow and Ice Data Center/World Data Center - A for Glaciology. Digital data available from nsidc@kryos.colorado.edu, University of Colorado, Boulder, Colorado.

Arnold, N., and M. Sharp, 1992: Influence of glacier hydrology on the dynamics of a large Quaternary ice sheet. *J. Quaternary Science*, **7**, 109-124.

Arpe, K., H. Behr, and L. Dümenil, 1997: Validation of the ECHAM4 climate model and re-analyses data in the Arctic region. *Proc. Workshop on the implementation of the Arctic Precipitation Data Archive at the Global Precipitation Climatology Centre*, WCRP-98, WMO/TD No. 804, 31-40.

Asrar, G., and J. Dozier, 1994: *EOS: Science Strategy for the Earth Observing System*. NASA, 119 pp.

Bales, R., R. Galarraga, and K. Elder, 1992: Distributed approach to modeling snowmelt runoff in alpine catchments. *Proc. Workshop on the Effects of Global Climate Change on Hydrology and Water Resources at the Catchment Scale*, Tsukuba, Japan, Feb. 3-6, 1992, 207-217.

Barnett, T. P., L. Dümenil, V. Schlese, E. Roeckner, and M. Latif, 1989: The effect of Eurasian snow cover on regional and global climate variations. *J. Atmos. Sci.*, **46**, 661-685.

Baron, J., L. E. Band, S. W. Running, and D. W. Cline, 1993: The effects of snow distribution on the hydrologic simulation of a high elevation Rocky Mountain watershed using Regional HydroEcological Simulation System, RHES Sys. *EOS. Transactions*, American Geophysical Union, Supplement, **74**, 237.

Barry, R. G., 1996: The parameterization of surface albedo for sea ice and its snow cover. *Progr. Phys. Geogr.*, **20**, 61-77.

Barry, R. G., 1997a: Satellite-derived data products for the polar regions. *EOS*, **78**(5), 52.

Barry, R. G., 1997b: Cryospheric data for model validations: Requirements and status. *Ann. Glaciol.*, in press.

Barry, R. G., and J. A. Maslanik. 1993: Monitoring lake freeze-up, breakup as a climatic index. In: R. G. Barry, B. E. Goodison and E.F. LeDrew (eds.), *Snow Watch '92*, Glaciology Data Report, GD-25. WDC-A for Glaciology, University of Colorado, Boulder, p. 66-79.

Barry, R. G., and A. M. Brennan, 1993: Towards a permafrost information system. In: *Permafrost, Sixth International Conference Proceedings, Vol. 1*. South China University of Technology Press, Wushan Guangzhou, China.

Barry, R. G., J. -M. Fallot, and R. L. Armstrong, 1995: Twentieth-century variability in snow cover conditions and approaches to detecting and monitoring changes: status and prospects. *Prog. in Phys. Geog.*, **19**, 520-532.

Basist, A., D. Garrett, R. Ferraro, N. Grody, and K. Mitchell, 1996: A comparison between snow cover products derived from visible and microwave satellite observations. *J. Appl. Met.*, **35**, 163-177.

Battisti, D. S., C. M. Bitz, and R. E. Moritz, 1997: Do general circulation models underestimate the natural variability in the Arctic climate? *J. Climate*, **10**, 1909-1920.

Belt, D., 1997: An Arctic Breakthrough. *National Geographic*, 191(2), 36-57.

Bentley, C. R., 1983: The west Antarctic ice sheet: diagnosis and prognosis. *Proc. Carbon Dioxide Research Conference: Carbon Dioxide, Science and Consensus.* (Berkeley Springs, W. VA., September 19-23, 1982).

Bentley, C. R., 1985: Glaciological evidence: the Ross Sea sector. Glaciers, ice sheets, and sea level: effect of a CO_2-induced climatic change. Report of a workshop held in Seattle, Washington September 13-15, 1984, United States Department of Energy, Washington, D.C., 178-196.

Bentley, C. R., and M. B. Giovinetto, 1991: Mass balance of Antarctica and sea level change. In: G. Weller, C. L. Wilson and B. A. B. Severin (eds.), *Polar regions and climate change*. University of Alaska, Fairbanks, p. 481-488.

Bernier, M., and J. P. Fortin, 1998: The potential of time series of C-band SAR data to monitor dry and shallow snow cover. *IEEE Transactions on Geoscience and Remote Sensing*, **36**, 1-18.

Bitz, C. M., D. S. Battisti, R. E. Moritz, and J. A. Beesley, 1996: Low-frequency variability in the arctic atmosphere, sea ice, and upper-ocean climate system. *J. Climate*, **9**, 394-408.

Bjørgo, E., O. M. Johannessen, and M. W. Miles, 1997: Analysis of merged SSMR-SSMI time series of Arctic and Antarctic sea ice parameters 1978-1995. *Geophys. Res. Lett.*, **24**, 413-416.

Blöschl, G., R. Kirnbauer, and D. Gutknecht, 1991a: Distributed snowmelt simulations in an Alpine catchment 1. Model evaluation on the basis of snow cover patterns. *Water Resour. Res.*, **27**(12), 3171-3179.

Blöschl, G., D. Gutknecht, and R. Kirnbauer, 1991b: Distributed snowmelt simulations in an Alpine catchment 2. Parameter study and model predictions. *Water Resour. Res.*, **27**(12), 3181-3188.

Bourke, R. H., and R. P. Garrett, 1987: Sea ice thickness distribution in the Arctic Ocean. *Cold Regions Science and Technology*, **13**, 259-280.

Braaten, R. O., 1997: *The Canadian snow water equivalent dataset.* Contract report prepared for Atmospheric Environment Service, Montreal, Canada, March 1997, 28 pp.

Broecker, W. S., 1994: Massive iceberg discharges as triggers for global climate change. *Nature*, **372**, 421-424.

Bromwich , D. H., R. -Y. Tzeng, and T. R. Parish, 1994: Simulation of the modern arctic climate by the NCAR CCMI. *J. Climate*, **7**, 1050-1069.

Bromwich, D. H., and B. Chen, 1995: Intercomparison of simulated polar climates by global climate models (Diagnostic Subproject 8). Abstracts of the First International AMIP Scientific Conference, Monterey, California, 33.

Brown, R. D., and P. Cote, 1992: Interannual variability in landfast ice thickness in the Canadian High Arctic, 1950-89. *Arctic*, **45**, 273-284.

Brown, R. D., and B. E. Goodison, 1996: Interannual variability in reconstructed Canadian snow cover, 1915-1992. *J. Climate*, **9**, 1299-1318.

Brown, R. D., 1997: Historical variability in Northern Hemisphere spring snow covered area. *Annals of Glaciology*, **25** (in press).

Brugman, M., and P. Raistrick and A. Pietroniro, 1997: Glacier related impact of doubling CO_2 on British Columbia and Yukon. In *Proc. of Workshop on "Responding to Global Climate Change in British Columbia and the Yukon"*, February 27-28, 1997, Vancouver, B.C., E. Taylor and W. Taylor (*eds.*), Environment Canada and British Columbia Ministry of Environment, Lands and Parks, pp. xx-xx [M. Brugman e-mailed 26/11].

Cavalieri, D. J., 1992: The validation of geophysical products using multisensor data. In *Microwave Remote Sensing of Sea Ice*, F.D. Carsey (ed.), Am. Geophys. Union, Washington, 233-242.

Cavalieri, D. J., and J. C. Comiso, 1997: *Algorithm theoretical basis document (ATBD) for the AMSR sea ice algorithm*. NASA Internal Report, 44 pp. [*http://wwwghcc.msfc.nasa.gov/AMSR/atbd-amsr-seaice.pdf*]

Cavalieri, D. J., P. Gloersen, C. L. Parkinson, J. C. Comiso, and H. J. Zwally, 1997: Observed hemispheric asymmetry in global sea ice changes. *Science*, **278**, 1104-1106.

Cess, R. D., and 32 others, 1991: Interpretation of snow-climate feedback as produced by 17 General Circulation Models. *Science*, **253**, 888-892.

Chahine, M. T., 1992: The hydrological cycle and its influence on climate. *Nature*, **359**, 373-380.

Chang, A. T. C., J. L. Foster, and D. K. Hall, 1987: Nimbus-7 derived global snow cover parameters. *Annals of Glaciology*, **9**, 39-44.

Chang, A. T. C., J. L. Foster, and A. Rango, 1991: Utilization of surface cover composition to improve the microwave determination of snow water equivalent in a mountainous basin. *Intl. J. Remote Sensing*, **12**, 2311-2319.

Chang, A. T. C, and A. Rango, 1996: Algorithm Theoretical Basis Document (ATBD) for the AMSR snow water equivalent algorithm. NASA Internal Report, Version 1.0, November 15, 1996. [*http://wwwghcc.msfc.nasa.gov/AMSR/snow_ATBD.html*]

Chen, Q., D. H. Bromwich, and L. Bai, 1997: Precipitation over Greenland retrieved by a dynamic method and its relation to cyclonic activity. *J. Climate*, **10**, 839-870.

Cihlar J. et al., 1997: GCOS/GTOS plan for terrestrial climate-related observations. Version 2.0, June 1997, GCOS-32, WMO/TD No. 796. [*http://www.wmo.ch/gcos/pub/topv2_1.html*]

Clark, P. U., 1994: Unstable behavior of the Laurentide ice sheet over deforming sediment and its implications for climate change. *Quaternary Research*, **41**, 19-25.

Clarke, G. K. C., U. Nitsan, and W. S. B. Paterson, 1977: Strain heating and creep instability in glaciers and ice sheets. *Reviews of Geophysics and Space Physics*, **15**, 235-247.

Clarke, G. K. C., S. G. Collins, and D. E. Thompson, 1984: Flow, thermal structure, and subglacial conditions of a surge-type glacier. *Canadian Journal of Earth Sciences*, **21**, 232-240.

Cline, D., 1997: Snow surface energy exchanges and snowmelt at a continental, midlatitude Alpine site. *Water Resour. Res.*, **33**, 689-701.

Cohen, J., and D. Rind, 1991: The effect of snow cover on the climate. *J. Climate*, **4**, 689-706.

Cogley, G. J., W. P. Adams, M. A. Ecclestone, F. Jung-Rothenhausler, and C. S. L. Ommanney, 1996: Mass balance of White Glacier, Axel Heiberg Island, N.W.T., Canada, 1960-91. *J. Glaciology*, **42**(142): 548-563.

Curry, J. A., J. L. Schramm, and E. E. Ebert, 1995: Sea ice-albedo climate feedback mechanism. *J. Climate*, **8**, 240-247.

Curry, J. A., W. B. Rossow, D. Randall, and J. E. Schramm, 1996: Overview of Arctic cloud and radiation characteristics. *J. Climate*, **9**, 1731-1764.

Cutler, P. M., and D. S. Munro, 1996: Visible and near-infrared reflectivity during the ablation period on Peyto Glacier, Alberta, Canada. *J. Glaciology*, **42**(141), 333-340.

de Sëve, D., M. Bernier, J.-P. Fortin, and A. Walker, 1997: Preliminary analysis of snow microwave radiometry using SSM/I passive microwave date: the case of the La Grand River watershed (Quebec). *Annals of Glaciology*, **25**, 353-361.

Dewey, K. F., and R. Heim, Jr., 1982: A digital archive of Northern Hemisphere snow cover, November 1966 through December 1980. *Bull. Amer. Met. Soc.*, **63**, 1132-1141.

Dozier, J., and J. Frew: Rapid calculation of terrain parameters for radiation modeling from digital elevation data. *IEEE Transactionson Geoscience and Remote Sensing*, **28**, 963-969.

Drewry D. J., 1982: Antarctica unveiled. *New Scientist*, **1315**, 246-251.

Duguay, C. R., and P. M. Lafleur, 1997: Monitoring ice freeze-up and break-up of shallow tundra lakes and ponds using ERS-1 SAR data. *Proceedings GER'97 - International Symposium: Geomatics in the Era of Radarsat*, May 24-30, 1997, Ottawa, Ontario (CD-ROM Vol. 1)

Duguay, C. R., D. W. Leverington, and H. McNairn, 1997: Land cover information content of polarimetric SAR data of a boreal forest, central Yukon Territory. *Proceedings GER '97- International Symposium: Geomatics in the Era of Radarsat*, May 24-30, 1997, Ottawa, Ontario (CD-ROM, Vol. 1).

Duguay, C. R., W. R. Rouse, P. M. Lafleur, D. L. Boudreau, Y. Crevier, and T. Pultz, 1998: Analysis of multi-temporal ERS-1 SAR data of subarctic tundra and forest in the northern Hudson Bay Lowland and implications for climate studies. *Canadian J. of Remote Sensing* (in review).

Easterling, D. R., P. Jamason, D. Bowman, P. Y. Hughes, and E. H. Mason, 1997: Daily snowdepth measurements from 195 stations in the United States. ORNL/CDIAC-95, NDP-059.

Ebert, E. E., and J. A. Curry. 1993: An intermediate one-dimensional thermodynamic sea ice model for investigating ice-atmosphere interactions. *J. Geophys. Res.*, **98**(C6), 10,085-10,109.

Elder, K., J. Dozier, and J. Michaelsen, 1991: Snow accumulation and distribution in an alpine watershed. *Water Resour. Res.*, **27**, 1541-1552.

Fabre, A., A. Letréguilly, C. Ritz, and A. Mangeney, 1994: Greenland under changing climates: sensitivity experiments with a new three-dimensional ice-sheet model. *Annals of Glaciology*, **21**, 1-7.

Fallot, J -M., R. G. Barry, and D. Hoogstrate, 1997: Variations of mean cold season temperature, precipitation and snow depth during the last 100 years in the former Soviet Union (FSU). *Hydrological Sciences*, **42**, 301-327.

Fang, X., and H. G. Stefan, 1996: Long-term lake water temperature and ice cover simulations/measurements. *Cold Regions Sci. and Technol.*, **24**, 289-304.

Fitzharris, B., 1996: The Cryosphere: Changes and their Impacts. In: *Climate Change 1995, Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses*, Cambridge University Press, 241-265.

Flato, G. M., and W. D. III. Hibler, 1992: Modeling pack ice as a cavitating fluid. *J. Physical Oceanography*, **22**, 626-651.

Flato, G. M., and W. D. III. Hibler, 1995: Ridging and strength in modeling the thickness distribution of Arctic sea ice. *J. Geophys. Res.*, **100**(C9), 18,611-18,626.

Flato, G. M., 1995: Spatial and temporal variability of Arctic ice thickness. *Annals of Glaciology*, **21**, 323-329.

Flato, G. M., and R. D. Brown, 1996: Variability and climate sensitivity of landfast Arctic sea ice. *J. Geophys. Res.*, **101**(C10), 25,767-25,777.

Flato, G. M., and D. Ramsden, 1997: Sensitivity of an atmospheric general circulation model to the parameterization of leads in sea-ice. *Annals of Glaciology*, **25** (in press).

Foster, D. J.Jr., and R. D. Davy, 1988: *Global snow depth climatology*. USAF publication USAFETAC/TN-88/006, Scott Air Force Base, Illinois, 48 pp.

Foster, J. L., D. K. Hall, A. T. C. Chang, and A. Rango, 1984: An overview of passive microwave snow research and results. *Reviews of Geophysics*, **22**, 195-208.

Foster, J., G. Liston, R. Koster, R.. Essery, H. Behr, L. Dumenil, D. Verseghy, S. Thompson, D. Pollard, and J. Cohen, 1996: Snow cover and snow mass intercomparisons of general circulation models and remotely sensed datasets. *J. Climate*, **9**, 409-426.

Frei, A., and D. A. Robinson, 1995: Northern Hemispheric snow cover extent: comparison of AMIP results to observations. *Proc. First Int'l. AMIP Scientific Conference*, Monterey, CA, U.S.A., 15-19 May, 1995. WMO TD No. 732, 499-504.

Gates, W. L., and 9 others, 1996: Climate Models - Evaluation. Chapter 5 in IPCC (1996).

Gloersen, P., W. J. Campbell, D. J. Cavalieri, J. C. Comiso, C. L. Parkinson, and H. J. Zwally, 1992: Arctic and Antarctic Sea Ice, 1978-1987: Satellite Passive-Microwave Observations and Analysis. NASA SP-511, National Aeronautics and Space Administration, Washington, D.C., 290 pp.

Gloersen, P., W. J. Campbell, D. J. Cavalieri, J. C. Comiso, C. L. Parkinson, and H. J. Zwally, 1993: Satellite passive microwave observations and analysis of Arctic and Antarctic sea ice, 1978-1987. *Annals of Glaciology*, **17**, 149-154.

Goita, K., A. E. Walker, B. E. Goodison, and A. T. C. Chang, 1997: Estimation of snow water equivalent in the boreal forest using passive microwave data. Proc. GER'97 (International Symposium: Geomatics in the Era of Radarsat), May 24-30, 1997, Ottawa, Ontario (CD-ROM Vol. 1).

Gold, L. W., and A. H. Lachenbruch, 1973: *Thermal conditions in permafrost—a review of North American literature*. Permafrost: North American contribution to the Second International Conference, Yakutsk, U.S.S.R., Washington, D.C., National Academy of Sciences, 3-25.

Goodison, B. E., 1989: Determination of areal snow water equivalent on the Canadian prairies using microwave radiometry. *Proc. IGARSS'89*, Vancouver, July 1989, **3**, 1243-1246.

Goodison, B. E., and A. E. Walker, 1994: Canadian development and use of snow cover information from passive microwave satellite data. *Proc. EAS/NASA International Workshop on Passive Microwave Remote Sensing Research Related to Land-Atmosphere Interactions*, St. Lary, France, January 11-15, 1993, 245-262.

Goodison, B. E., and R. D. Brown, 1997: CRYSYS - Use of the cryosphere system for monitoring global change in Canada. *Proc. GER'97* (International Symposium: Geomatics in the Era of Radarsat), May 24-30, 1997, Ottawa, Ontario (CD-ROM Vol. 1).

Gratton, D. J., P. J. Howarth, and D. J. Marceau, 1994: An investigation of terrain irradiance in a mountain glacier basin. *J. Glaciology*, **40**(136): 519-526.

Groisman, P. Ya, T. R. Karl, and R. W. Knight, 1994a: Observed impact of snow cover on the heat balance and the rise of continental spring temperatures. *Science*, **263**, 198-200.

Groisman, P. Ya, T. R. Karl, and R. W. Knight, 1994b: Changes of snow cover, temperature and radiative heat balance over the Northern Hemisphere. *J. Climate*, **7**, 1633-1656.

Groisman, P. Ya, and D. R. Easterling, 1994: Variability and trends of total precipitation and snowfall over the United States and Canada. *J. Climate*, **7**, 184-205.

Gutzler, D. S., and J. W. Preston, 1997: Evidence for a relationship between spring snow cover in North America and summer rainfall in New Mexico. *Geophys. Res. Lett.*, **24**, 2207-2210.

Haeberli, W., D. Barsch, A. E. Corte, A. P. Gorbunov, S. A. Harris, M. Hoelzle, L. King, G. K. Lieb, R. S. Oedegard, J. L. Sollid, D. Trombotto, and D. Vonder Mühll, 1995: Monitoring of mountain permafrost: a review of ongoing programmes. International Permafrost Association workshop 'Our Current Understanding of Processes and Ability to Detect Global Change', Hanover, New Hampshire, 9-11 December 1995.

Haeberli, W., 1995: Glacier fluctuations and climate change detection - operational elements of a worldwide monitoring strategy. WMO Bulletin, **44**, 23-31.

Halekinen, S., 1993: An Arctic source for the Great Salinity Anomaly: A simulation of the Arctic ice-ocean system. *J. Geophys. Res.*, **98**, 16,397-16,410.

Hall, D. K., and J. Martinec, 1985: *Remote sensing of snow and ice*. Chapman and Hall, New York, USA, 189 pp.

Hall, D. K., 1993: Active and passive microwave remote sensing of frozen lakes for regional climate studies. *SNOW WATCH '92 - Detection Strategies for Snow and Ice*. World Data Center A for Glaciology, Glaciological Data Report GD-25, 80-85

Hall, D. K., D. B. Fagre, F. Klasner, G. Linebaugh, and G. E. Liston, 1994: Analysis of ERS 1 synthetic aperture radar data of frozen lakes in northern Montana and implications for climate studies. *J. Geophys. Res.*, **99**(C11), 22,473-22,482.

Hall, D. K., G. A. Riggs, and V. V. Salomonson, 1995: Development of methods for mapping global snow cover using moderate resolution imaging spectroradiometer data. *Rem. Sens. of Environ.*, **54**, 127-140.

Hall, D. K., and 5 others, 1996: Algorithm Theoretical Basis Document (ATBD) for the MODIS snow-, lake ice- and sea ice-mapping algorithms. NASA Internal Report, Version 3.0, November 1, 1996. [*http://modarch.gsfc.nasa.gov/MODIS/ATBD/atbd_mod10.pdf*].

Hall, D. K., 1996: Remote sensing applications to hydrology: imaging radar. *Hydrological Sciences*, **41**, 609-624.

Harbor, J., M. Sharp, L. Copland, B. Hubbard, P. Nienow, and D. Mair, 1997: Influence of subglacial drainage conditions on the velocity distribution within a glacier cross section. *Geology*, **25**, 739-742.

Harder, M., 1997: Role of precipitation in numerical simulations of arctic sea ice and related freshwater balance. *Proc. Workshop on the implementation of the Arctic Precipitation Data Archive at the Global Precipitation Climatology Centre*, WCRP-98, WMO/TD No. 804, 26-30.

Hartman, R. K., A. A. Rost, and D. M. Anderson, 1995: Operational processing of multi-source snow data. *Proc. Western Snow Conference*, 147-151.

Heron, R., and M -K. Woo, 1994: Decay of a high Arctic lake-ice cover: observations and modeling. *J. Glaciology*, **40**, 283-292.

Hibler, W. D. III, and J. E. Walsh, 1982: On modeling seasonal and interannual fluctuations of Arctic sea ice. *J. Physical Oceanography*, **12**, 1514-1523.

Hibler, W. D. III., 1979: A dynamic thermodynamic sea ice model. *J. Physical Oceanography*, **9**, 817-846.

Hibler, W. D. III., 1984: *Ice dynamics. Sea ice, dynamics, pressure ridges.* U.S. Army CRREL, Monograph 84-3, Vol. 3, Hanover, N.H., U.S. Army CRREL, 52 pp.

Hughes, M. G., and D. A. Robinson, 1996: Historical snow cover variability in the Great Plains region of the USA: 1910 to 1993. *Intl. J. Climatology*, **16**, 1005-1018.

Hughes, M. G., A. Frei, and D. A. Robinson, 1996: Historical analysis of North American snow cover extent: merging satellite and station-derived snow cover observations. *Proc. 53rd Eastern Snow Conference*, Williamsburg, Virginia, 21-31.

Huybrechts, P., 1990: The Antarctic ice sheet during the last glacial-interglacial cycle: a three-dimensional experiment. *Annals of Glaciology*, **14**, 115-119.

Huybrechts, P., A. Letréguilly, and N. Reeh, 1991: The Greenland Ice Sheet and greenhouse warming. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **89**, 399-412.

Huybrechts, P., 1993: Glaciological modeling of the late Cenozoic East Antarctic ice sheet: stability or dynamism? *Geografiska Annaler*, **75A**, 221-238.

Huybrechts, P., and J. Oerlemans, 1988: Evolution of the East Antarctic ice sheet: a numerical study of thermo-mechanical response patterns with changing climate. *Annals of Glaciology*, **11**, 52-59.

Iken, A., 1981: The effect of the subglacial water pressure on the sliding velocity of a glacier in an idealised numerical model. *J. Glaciology*, **27**, 407-421.

Iken, A., and R. A. Bindschadler, 1986: Combined measurement of subglacial water pressure and surface velocity of Findelengletscher, Switzerland: conclusions about drainage system and sliding mechanism. *J. Glaciology*, **32**, 101-119.

Imbrie, J., and 18 others, 1993: On the structure and origin of major glaciation cycles: 2. The 100,000-year cycle. *Paleoceanography*, **8**, 699-735.

Ip, C. F., W. D.III. Hibler, and G. M. Flato, 1991: On the effect of rheology on seasonal sea-ice simulations. *Annals of Glaciology*, **15**, 17-25.

Ip, C. F., 1993: *Numerical investigation of different rheologies on sea ice dynamics*. Ph.D. thesis, Thayer School of Engineering, Dartmouth College, Hanover, N.H., 242 pp.

IPCC, 1990: *Climate Change - The IPCC Scientific Assessment*. J. T. Houghton, G. J. Jenkins and J. J. Ephraums (eds.), Cambridge University Press, Cambridge, UK, 365 pp.

IPCC, 1996: *Climate Change 1995: The Science of Climate Change*. Houghton, J. T., L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell (eds.), Contribution of WGI to the Second Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, UK, 572 pp.

Jacobs, S. S., H. H. Helmer, C. S. M. Doake, A. Jenkins, and R. M. Frohlich, 1992: Melting of ice shelves and the mass balance of Antarctica. *J. Glaciology*, **38**, 375-387.

Jeffries, M. O., K. Morris, and G. E. Liston, 1996: A method to determine lake depth and water availability on the north slope of Alaska with spaceborne imaging radar and numerical ice growth modeling. *Arctic*, **49**(4), 367-374.

Jeffries, M. O., K. Morris, W. F. Weeks, and H. Wakabayashi, 1994: Structural and stratigraphic features and ERS 1 synthetic radar backscatter characteristics of ice growing on shallow lakes in NW Alaska, winter 1992-1992. *J. Geophys. Res.*, **99**(C11), 22,459-22,471.

Johannessen, M., and A. Henriksen, 1978: Chemistry of snowmelt: changes in concentration during melting. *Water Resour. Res.*, **14**, 615-619.

Johannessen, O. M., M. Miles, and E. Bjørgo, 1995: The Arctic's shrinking sea ice. *Nature*, **376**, 126-127.

Joughin, I. R., D. P. Winebrenner, and M. A. Fahnestock, 1995: Observations of ice-sheet motion in Greenland using satellite radar interferometry. *Geophys. Res. Lett.*, **22**, 571-574.

Kamb, B., 1987: Glacier surge mechanism based on linked cavity configuration of the basal water conduit system. *J. Geophys. Res.*, **92**, 9083-9100.

Kamb, B., and H. Engelhardt, 1991: Antarctic ice stream B: conditions controlling its motions and interactions with the climate system. *IAHS Publication 208*, 145-154.

Kane, D. L., L. D. Hinzman, and J. P. Zarling, 1991: Thermal response of the active layer to climatic warming in a permafrost environment. *Cold Regions Science and Technology*, **19**, 111-122.

Kite, G. W., 1991: A watershed model using satellite data applied to a mountain basin in Canada. *J. Hydrol.*, **128**(1-4), 157-169.

Knox, J. C., 1993: Large increases in flood magnitude in response to modest changes in climate. *Nature*, **361**, 430-432.

Koch, E., and E. Rüdel, 1990: Mögliche Auswirkungen eines verstärkte Treibhauseffekte auf die Schneeverhältnisse in Österreich. *Wetter und Leben*, **45**, 137-153.

Koerner, R. M., 1989: Ice core evidence for extensive melting of the Greenland Ice Sheet in the last interglacial. *Science*, **244**, 964-968.

Koster, E. A., 1993: Introduction - Present global change and permafrost, within the framework of the International Geosphere-Biosphere Programme. *Permafrost and Periglacial Processes*, **4**, 95-98.

Kuusisto, E., 1993: Lake ice observations in Finland in the 19th and 20th Century: any message for the 21st? In: R.G. Barry, B.E. Goodison and E.F. LeDrew (*eds.*), *Snow Watch '92*, Glaciology Data Report, GD-25. WDC-A for Glaciology, University of Colorado, Boulder, 57-65.

Kwok, R., D. A. Rothrock, H. L. Stern, and G. F. Cunningham, 1995: Determination of the age distribution of sea ice from Lagrangian observations of ice motion. *IEEE Transactions on Geoscience and Remote Sensing*, **33**, 392-400.

Lachenbruch, A. H., J. H. Sass, B. V. Marshall, and T. H. Jr. Moses, 1982: Permafrost, heat flow, and the geothermal regime at Prudhoe Bay, Alaska. *J. Geophys. Res.*, **87**(B11), 9301-9316.

Lachenbruch, A. H., and B. V. Marshall, 1986: Changing climate - geothermal evidence from permafrost in the Alaskan arctic. *Science*, **234**, 689-696.

Lavoie, C., and S. Payette, 1992: Black Spruce growth forms as a record of changing winter environment at treeline, Quebec, Canada. *Arctic and Alpine Res.*, **24**, 40-49.

Lawby, C. P., D. J. Smith, C. P. Laroque, and M. M. Brugman, 1995: Glaciological studies at Rae Glacier, Canadian Rockies. *Physical Geography*, **15**(5), 425-441.

Leathers, D. J., and D. A. Robinson, 1993: The association between extremes in North American snow cover extent and United States temperatures. *J. Climate*, **6**, 1345-1355.

Ledley, T. S., 1991: Snow on sea ice: competing effects in shaping climate. J. Geophys. Res., 96, 17,195-17,208.

Ledley, T. S., 1993: Variations in snow on sea ice: a mechanism for producing climate variations. *J. Geophys. Res.*, **98**(D6), 10,401-10,410.

Lemke, P., W. B. Owens, and W. D. III. Hibler, 1990: A coupled sea-ice mixed-layer pycnocline model for the Weddell Sea. *J. Geophys. Res.*, **95**, 9513-9525.

Lemke, P., L. Anderson, R. Barry, and V. Vuglinsky (*eds.*), 1996: *Proceedings of the ACSYS Conference on the Dynamics of the Arctic Climate System*, Goteborg, Sweden, 7-10 November 1994, WCRP-94, WMO/TD 602, no. 760, 483 pp.

Letréguilly, A., P. Huybrechts, and N. Reeh, 1991: Steady-state characteristics of the Greenland ice sheet under different climates. *J. Glaciology*, **37**, 149-157.

Leverington, D. W., and C. R. Duguay, 1996. Evaluation of three supervised classifiers in mapping “depth to late-summer frozen ground”, central Yukon Territory. *Canadian J. Remote Sensing*, **22**, 163-174.

Leverington, D. W., and C. R. Duguay, 1997: A neural network method to determine the presence or absence of permafrost near Mayo, Yukon Territory, Canada. *Permafrost and Periglacial Processes*, **8**, 205-215.

Lewkowicz, A. G., and C. R. Duguay, 1995: Assessment of SPOT panchromatic imagery in the detection and identification of permafrost features, Fosheim Peninsula, Ellesmere Island, N.W.T. *Proceedings of the 17th Canadian Symposium on Remote Sensing*, Saskatoon, Saskatchewan, June 13-15, pp. 8-14.

Lindsay, R. W., and D. A. Rothrock, 1994: Arctic sea ice albedo from AVHRR. *J. Climate*, **7**, 1737-1749.

Liu, A. K., S. Martin, and R. Kwok, 1997: Tracking of ice edges and ice floes by wavelet analysis of SAR images. *J. Atmos. Ocean. Tech.*, **14**, 1187-1198.

Loth, B., H. -F. Graf, and J. M. Oberhuber. 1993: Snow cover model for global climate simulations. *J. Geophys. Res.*, **98**(D6), 10,451-10,464.

Lynch-Stieglitz, M., 1994: The development and validation of a simple snow model for the GISS GCM. *J. Climate*, **7**, 1842-1855.

MacAyeal, D. R., 1992: Irregular oscillations of the West Antarctic ice sheet. *Nature*, **359**, 29-32.

MacAyeal, D. R., 1993a: A low-order model of growth/purge oscillations of the Laurentide Ice Sheet. *Paleoceanography*, **8**, 767-773.

MacAyeal, D. R., 1993b: Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic’s Heinrich events. *Paleoceanography*, **8**, 775-784.

Marks, D., and J. Dozier, 1992: Climate and energy exchange at the snow surface in the alpine region of the Sierra Nevada. 2. Snow cover energy balance. *Water Resour. Res.*, **28**, 3043-3054.

Markus, T., and D. J. Cavalieri, 1997: Snow depth distribution over sea ice in the Southern Ocean from satellite passive microwave data. American Geophysical Union Antarctic Research Series. M. Jeffries (*ed.*). (in press)

Marsh, P., 1991: Water flux in melting snow covers. In: *Advances in Porous Media, Vol. 1*. M.Y. Corapcioglu (ed.), Elsevier, Amsterdam, 61-124.

Marshall, S., J. O. Roads, and G. Glatzmaier, 1994: Snow hydrology in a general circulation model. *J. Climate*, **7**, 1251-1269.

Martin, S., K. Steffen, J. Comiso, D. Cavalieri, M. R. Drinkwater, and B. Holt, 1992: Microwave remote sensing of polynyas. In: Carsey, F. D. (*ed.*), Microwave remote sensing of sea ice, Washington, DC, American Geophysical Union, 1992, 303-311.

Martinec, J., and A. Rango, 1991: Indirect evaluation of snow reserves in mountain basins. In: H. Bergmann, H. Lang, W. Frey, D. Issler, and B. Salm (eds.), International Association of Hydrological Sciences. IAHS/AISH Publication 602, no.205. *Proc. International Symposium on Snow, Hydrology and Forests in High Alpine Areas*, Vienna, 11-14 August 1991, 111-119.

Martinec, J., K. Seidel, U. Burkart, R. Baumann, H. Bergmann, H. Lang, W. Frey, D. Issler, and B. Salm (*eds.*), 1991: Areal modeling of snow water equivalent based on remote sensing techniques. Snow, hydrology and forests in high alpine areas. Proceedings of an international symposium held during the 20th General Assembly of the International Union of Geodesy and Geophysics at Vienna, 11-24 August 1991, IAHS Press, Wallingford, Oxfordshire, 121-129.

Maslanik, J. A., and R. G. Barry, 1987: Lake ice formation and breakup as an indicator of climate change: potential for monitoring using remote sensing techniques. In: S.I. Solomon, M. Beran, and W. Hogg (*eds.*), Influence of Climate Change and Climatic Variability on the Hydrologic Regime and Water Resources, Proceedings of the Vancouver Symposium, IAHS Publication no. 168, 153-161.

Matson, M., 1991: NOAA satellite snow cover data. *Palaeogeography, Palaeoclimatology, Palaeoecology* (Global and Planetary Change Section), **90**(1-3), 213-218.

Matson, M., C. F. Ropelewski, and M. S. Varnadore, 1986: *An atlas of satellite-derived Northern Hemisphere snow cover frequency*, National Weather Service, Wash., D.C., 75 pp.

Maykut, G. A. 1982: Large-scale heat exchange and ice production in the central Arctic. *J. Geophys. Res.*, **87**(C10), 7971-7984.

Maykut, G. A., 1978: Energy exchange over young sea ice in the central Arctic. *J. Geophys. Res*., **83**(C7), 3646-3658.

McGinnis, D. L., and R. G. Crane, 1994: A multivariate analysis of Arctic climate in GCMs. *J. Climate*, **7**, 1240-1250.

McLaren, A. S., J. S. Walsh, R. H. Bourke, R. L. Weaver, and W. Wittmann, 1992: Variability in sea-ice thickness over the North Pole from 1977 to 1990. *Nature*, **358**, 224-226.

Meehl, G. A., 1994: Influence of the land surface in the Asian summer monsoon: external conditions versus internal feedbacks. *J. Climate*, **7**, 1033-1049.

Meier, M. F., 1983: Snow and ice in a changing hydrological world. *Hydrological Sciences Journal*, **28**, 3-22.

Meier, M. F., 1984: Contribution of small glaciers to global sea level rise. *Science*, **226**, 1418-1421.

Mercer, J. H., 1968: *Antarctic ice and Sangamon sea level*. International Association of Hydrological Sciences, Publication No. 179, 217-225.

Morassutti, M. P., 1989: Surface albedo parameterization in sea-ice models. *Prog. Physical Geography*, **13**, 348-366.

Morris, K., M. O. Jeffries, and W. F. Weeks, 1995: Ice processes and growth history on arctic and sub-arctic lakes using ERS-1 SAR data, *Polar Record*, **31**(177), 115-128.

Morrissey, L. A., L. Strong, and D. H. Card, 1986: Mapping permafrost in the boreal forest with thematic mapper satellite data. *Photogrammetric Engineering and Remote Sensing*, **52**, 1513-1520.

Moritz, R. E., J. A. Curry, A. S. Thorndike, and N. Untersteiner, 1993: *SHEBA—a research program on the Surface Heat Budget of the Arctic Ocean*. Arctic System Science, Ocean-Atmosphere-Ice Interactions (ARCSS OAII) Science Management Office, 34 pp.

Munro, D. S., 1990: Comparison of melt energy computations and ablatometer measurements on melting ice and snow. *Arctic and Alpine Research*, **22**, 153-162.

Munro, D. S., and G. J. Young, 1982: An operational net shortwave radiation model for glacier basins. *Water Resour. Res.*, **18**(2), 220-230.

Oerlemans, J., 1986: Glaciers as indicators of a carbon dioxide warming. *Nature*, **320**, 607-609.

Oerlemans, J., 1994: Quantifying global warming from the retreat of glaciers. *Science*, **264**, 243-245.

Ohmura, A., M. Wild, and L. Bengtsson, 1996: A possible change in mass balance of the Greenland and Antarctic ice sheets in the coming century. *J. Climate*, **9**, 2124-2135.

Oke, T. R., 1987: *Boundary Layer Climates*. Second Edition, Routledge, 435 pp.

Osterkamp, T. E. 1984. Potential impact of a warmer climate on permafrost in Alaska. In: McBeath, J.H. (*ed.*), Potential effects of carbon dioxide-induced climatic changes in Alaska, The proceedings of a conference. Fairbanks, University of Alaska, March 1984, 106-113.

Osterkamp, T. E., 1994: Evidence for warming and thawing of discontinuous permafrost in Alaska. *EOS*, **75**, 85.

Palecki, M. A., and R. G. Barry. 1986: Freeze-up and break-up of lakes as an index of temperature changes during the transition seasons: A case study in Finland. *J. Climate and Appl. Meteorology*, **25**, 893-902.

Parkinson, C. L., 1995: Recent sea-ice advances in Baffin Bay/Davis Strait and retreats in the Bellinshausen Sea. *Annals of Glaciology*, **21**, 348-352.

Parkinson, C. L., and W. M. Washington, 1979: A large-scale numerical model of sea ice, *J. Geophys. Res.*, **84**(C1), 311-337.

Parkinson, C. L., J. C. Comiso, H. J. Zwally, D. J. Cavalieri, P. Gloersen, and W. J. Campbell, 1987: Arctic Sea Ice, 1973-1976: Satellite Passive-Microwave Observations, NASA SP-489, National Aeronautics and Space Administration, Washington, D.C., 296 pp.

Paterson, W. S. B., 1994: *The Physics of Glaciers*, Third Edition. Pergamon Press, Oxford.

Paterson, W. S. B., 1993: World sea level and the present mass balance of the Antarctic ice sheet. In: W.R. Peltier (*ed.*), *Ice in the Climate System*, NATO ASI Series, I12, Springer-Verlag, Berlin, 131-140.

Peddle, D. R., G. M. Foody, A. Zhang, S. E. Franklin, and E. F. LeDrew, 1994: Multi-source image classification II: An empirical comparison of evidential reasoning and neural network approaches. *Can. J. Remote Sensing*, **20**, 396-407.

Pelto, M. S., 1996: Annual net balance of North Cascade Glaciers, 1984-94. *J. Glaciology*, 42, 3-9.

Pollard, D., and S. L. Thompson, 1994: Sea-ice dynamics and CO_2 sensitivity in a global climate model. *Atmosphere-Ocean*, **32**(2), 449-467.

Pomeroy, J. W., P. Marsh, and D. M. Gray, 1997: Application of a distributed blowing snow model to the Arctic. *Hydrological Processes*, **11**, 1451-1464.

Prinsenberg, S. J. 1988: Ice-cover and ice-ridge contributions to the freshwater contents of Hudson Bay and Foxe Basin. *Arctic*, **41**, 6-11.

Prowse, T. D., 1990: Northern hydrology: an overview. In: Northern Hydrology: Canadian Perspectives, T. D. Prowse and C. S. L. Ommanney (eds.), *NHRI Science Report No. 1*, National Hydrology Research Institute, Environment Canada, Saskatoon, Saskatchewan, 1-36.

Randall, D. A., and 26 others, 1994: Analysis of snow feedbacks in 14 general circulation models. *J. Geophys. Res.*, **99**(D10), 20,757-20,771.

Rango, A., 1993: Snow hydrology processes and remote sensing. *Hydrological Processes*, **7**, 121-138.

Reeh, N., 1989: *Dynamic and climatic history of the Greenland Ice Sheet*. Chapter 14 in Quaternary Geology of Canada and Greenland, R.J. Fulton (*ed.*); Geological Survey of Canada, Geology of Canada, no. 1, 795-822.

Reycraft, J., and W. Skinner, 1993: Canadian lake ice conditions: an indicator of climate variability. *Climatic Perspectives*, **15**, 9-15.

Rignot, E., and J. Way, 1994. "Monitoring freeze-thaw cycles along north-south Alaskan transects using ERS-1 SAR." *Remote Sensing of Environment*, **49**, 131-137.

Rignot, E., K. C. Jezek, and H. G. Sohn, 1995: Ice flow dynamics of the Greenland Ice Sheet from SAR. *Geophysical Res. Lett.*, **22**, 575-578.

Rind, D., R. Healy, C. Parkinson, and D. Martinson, 1995: The role of sea ice in 2 x CO2 climate model sensitivity. Part I: The total influence of sea ice thickness and extent. *J. Climate*, **8**, 449-463.

Robinson, D. A., K. F. Dewey, and R. R. Heim, 1993: Global snow cover monitoring: an update. *Bull. Amer. Meteorol. Soc.*, **74**, 1689-1696.

Rosenthal, W., and J. Dozier, 1996: Automated mapping of montane snow cover at subpixel resolution from the LANDSAT Thematic Mapper. *Water Resour. Res.*, **32**, 115-130.

Rouse, W. R., and 9 others, 1997: Effects of climate changes on the fresh waters of Arctic and subarctic North America. *Hydrological Processes*, **11**, 873-902.

Scharfen, G., R. G. Barry, D. A. Robinson, G. Kukla, and M. C. Serreze, 1987: Large-scale patterns of snow melt on Arctic sea ice mapped from meteorological satellite imagery. *Annals of Glaciology*, **9**, 1-6.

Sellman, P. V., W. F. Weeks, and W. J. Campbell, 1975: *Use of side-looking airborne radar to determine lake depth on the Alaskan North Slope*. Special Report No. 230. Hanover, New Hampshire: Cold Regions Research and Engineering Laboratory.

Shi, J. C., J. Dozier, and R. E. Davis, 1990: Simulation of snow-depth estimation from multi-frequency radar. Proc. IGARSS 1990, IEEE No. 90CH2825-8, 1129-1132.

Shine, K. P., and R. G. Crane, 1984: The sensitivity of a one-dimensional thermodynamic sea ice model to changes in cloudiness. *J. Geophys. Res.*, **89**, 10,615-10,622.

Steffen, K., and A. Ohmura, 1985: Heat exchange and surface conditions in North Water, northern Baffin Bay. *Annals of Glaciology*, **6**, 178-181.

Stern, H. L., D. A. Rothrock, and R. Kwok, 1995: Open water production in Arctic sea ice: Satellite measurements and model parameterizations. *J. Geophys. Res.*, **100**(C10), 20,601-20,612.

Stieglitz, M., D. Rind, J. Famiglietti, and C. Rosenzweig, 1997: An efficient approach to modeling the topographic control of surface hydrology for regional and global climate modeling. *J. Climate*, **10**, 118-137.

Sturm, M., and J. B. Johnson, 1991: Natural convection in the subarctic snow cover. *J. Geophys. Res.*, **96**(B7), 11,657-11,671.

Sturm, M., J. Holmgren, and G. E. Liston. 1995: A seasonal snow cover classification system for local to global applications. *J. Climate*, **8**, 1261-1283.

Tao, X., J. Walsh, and W. Chapman, 1996: An assessment of global climate model simulations of arctic air temperatures. *J. Climate*, **9**, 1060-1076.

Tarboton, D. G., M. J. Al-Adhami, and D. S. Bowles, 1991: A preliminary comparison of snowmelt models for erosion prediction. *Proc. 59th Western Snow Conference*,79-90.

Thompson, S. L., and D. Pollard, 1997: Greenland and Antarctic mass balances for present and doubled atmospheric CO_2 from the GENESIS Version-2 global climate model. *J. Climate*, **10**, 871-900.

Thorndike, A. S., D. A. Rothrock, G. A. Maykut, and R. Colony, 1975: The thickness distribution of sea ice. *J. Geophys. Res.*, **80**, 4501-4513.

Titus, J. G., and V. K. Narayanan. 1995: The probability of sea level rise. United States Environmental Protection Agency, EPA 230-R-95-008.

Van den Broeke, M. R., 1996: The atmospheric boundary layer over ice sheets and glaciers. Utrecht, Universitiet Utrecht, 178 pp.

Van den Broeke, M. R., and R. Bintanja, 1995: The interaction of katabatic wind and the formation of blue ice areas in East Antarctica. *J. Glaciology*, **41**, 395-407.

Van de Wall, R. S. W., J. Oerlemans, and J. C. van der Hage, 1992: A study of ablation variations on the tongue of Hintereisferner, Austrian Alps. *J. Glaciology*, **38**, 319-324.

Vance, R. E., R. W. Mathewes, and J. J. Clague, 1992: 7000 year record of lake-level change on the northern Great Plains: a high resolution proxy of past climate. *Geology*, **20**, 879-882.

Vachon, P. W., D. Geudtner, K. Mattar, A. L. Gray, M. Brugman, and I. Cumming, 1996: Differential interferometry measurements of Athabasca and Saskatchewan glacier flow rate. *Can. J. Remote Sensing*, **22**, 287-296.

Verbitsky, M. Ya., and B. Saltzman, 1995: Behavior of the East Antarctic ice sheet as deduced from a coupled GCM/ice-sheet model. *Geophys. Res. Lett.*, **22**, 2913-2916.

Vernekar, A. D., J. Zhou, and J. Shukla, 1995: The effect of Eurasian snow cover on the Indian monsoon. *J. Climate*, **8**, 248-266.

Verseghy, D., 1991: CLASS A Canadian land surface scheme for GCMS. I: Soil model. *Intl. J. Climatology*, **11**, 111-133.

Walker, A. E., and M. R. Davey, 1993: Observation of Great Slave Lake ice freeze-up and break-up processes using passive microwave satellite data. *Proc. 16th Canadian Symp. on Remote Sensing*, Sherbrooke, Quebec, 7-10 June, 1993, 233-238.

Walker, A. E., and B. E. Goodison, 1993: Discrimination of wet snow cover using passive microwave satellite data. *Annals of Glaciology*, **17**, 307-311.

Walland, D. J., and I. Simmonds, 1996: Sub-grid-scale topography and the simulation of Northern Hemisphere snow cover. *Intl. J. Climatology*, **16**, 961-982.

Walsh, J. E., 1978: *A data set of northern hemisphere sea ice extent, 1953-76*. Glaciological Data. Report GD-2, National Snow and Ice Data Center, University of Colorado, Boulder, CO., 49-51.

Walsh, J. E., and W. L. Chapman, 1990: Arctic contribution to upper-ocean variability in the North Atlantic. *J. Climate*, **3**, 1462-1473.

Walsh, J. E., and R. G. Crane, R., 1992: A comparison of GCM simulations of Arctic climate. *Geophys. Res. Lett.*, **19(1)**, 29-32.

Walsh, J. E., 1995: Continental snow cover and climate variability. In: *Natural Climate Variability on Decadal-to-Century Time Scales*. National Academy Press, Washington, D.C., 49-58.

Washburn, A. L., 1973: *Periglacial processes and environments*. Edward Arnold, London, 320 pp.

WCRP, 1994: Initial implementation plan for the Arctic Climate System Study (ACSYS). WCRP-85 (WMO/TD-No. 627), Geneva.

WCRP, 1997: GCOS/GTOS Plan for Terrestrial Climate-Related Observations. Version 2.0. *GCOS*, **32** (WMO/TD No. 796), Geneva.

Weaver, A. J., and T. M. C. Hughes, 1994: Rapid interglacial climate fluctuations driven by North Atlantic ocean circulation. *Nature*, **367**, 447-450.

Weertman, J., 1972: General theory of water flow at the base of a glacier or ice sheet. *Reviews of Geophysics and Space Physics*, **10**, 287-333.

Welch, H. E., 1992: Energy flow through the marine ecosystem of the Lancaster Sound region, Arctic Canada. *Arctic*, **45**, 343.

Whillans, I. M., 1981: Reaction of the accumulation zone portions of glaciers to climatic change. *J. Geophys. Res.*, **86**(C5), 4274-4282.

Whillans, I. M., M. Jackson, and Y -H. Tseng, 1993: Velocity pattern in a transect across ice stream B, Antarctica. *J. Glaciology*, **39**, 562-572.

Williams, G. P., 1971: Predicting the date of lake ice breakup. *Water Resour. Res.*, **7**, 323-333.

Wolford, R. A., R. C. Bales, and S. Sorooshian, 1996: Development of a hydrochemical model for seasonally snow-covered alpine watersheds: application to Emerald Lake watershed, Sierra Nevada, California. *Water Resour. Res.*, **32**, 1061-1074.

Woo, M -K., and P. Steer, 1986: Monte Carlo simulation of snow depth in a forest. *Water. Resour. Res.*, **22**, 864-868.

Wynne, R. H., M. K. Clayton, T. M. Lillesand, and D. C. Rodman, 1996: Determinants of temporal coherence in the satellite derived 1978-1994 ice breakup dates of lakes on the Laurentian Shield. *Limnology and Oceanography*, **41**, 832-838.

Xu, H., J. O. Bailey, E. C. Barrett, and R. E. J. Kelly, 1993: Monitoring snow area and depth with integration of remote sensing and GIS. *Intl. J. Remote Sensing*, **14**, 3259-3268.

Yang, Z -L., R. E. Dickinson, A. Robock, and K. Ya. Vinnikov, 1997: Validation of the snow submodel of the Biosphere-Atmosphere Transfer Scheme with Russian snow cover and meteorological observational data. *J. Climate*, **10**, 353-373.

Young, G. J., M. C. English, and C. Hopkinson, 1996: *Changes in glacier dimensions 1951-1994 in the Bow River Basin above Banff.* CRYSYS Collaborative Research Agreement Report to Atmospheric Environment Service, Cold Regions Research Centre, Wilfrid Laurier University, 35 pp.

Zhang, T., T. E. Osterkamp, and K. Stamnes, 1996: Influence of the depth hoar layer of the seasonal snow cover on the ground thermal regime. *Water Resour. Res.*, **32**, 2075-2086.

Zuo, Z., and J. Oerlemans, 1996: Modeling albedo and specific balance of the Greenland Ice Sheet: calculations for the Sondre Stromfjord transect. *J. Glaciology*, **42**(141), 305-317.

Zwally, H. J., J. C. Comiso, C. L. Parkinson, W. J. Campbell, F. D. Carsey, and P. Gloersen, 1983: *Antarctic Sea Ice, 1973-1976: Satellite Passive-Microwave Observations*. NASA SP-459, National Aeronautics and Space Administration, Washington, D.C., 206 pp.

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