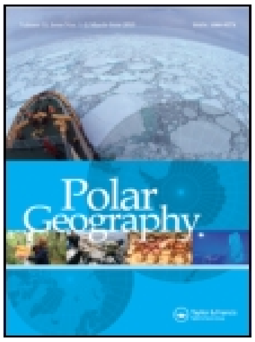

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THE ROLE OF SNOW AND ICE IN THE GLOBAL CLIMATE SYSTEM: A REVIEW[1]

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Abstract: Global snow and ice cover (the "cryosphere") plays a major role in global climate and hydrology through a range of complex interactions and feedbacks, the best known of which is the ice-albedo feedback. Snow and ice cover undergo marked seasonal and long-term changes in extent and thickness. In the Proterozoic era, for example, a long-lived "snowball" Earth has been proposed, while in the Pleistocene epoch, glacial and interglacial intervals alternated in a quasi-periodic manner, but with a smaller spatial extent. The perennial elements of the cryosphere—the major ice sheets and permafrost—play a role in present-day regional and local climate and hydrology, but the large seasonal variations in snow cover and sea ice are of importance on continental to hemispheric scales. The characteristics of these variations, especially in the Northern Hemisphere, and evidence for recent trends in snow and ice extent are discussed. The relative roles of natural variability in the climate-system forcing of such trends, versus possible anthropogenic influences, cannot yet be confidently separated. However, continued careful monitoring and assessment of the likely causes and their possible consequences of such changes is clearly a vital task for scientists studying climate-cryosphere processes. The World Climate Research Programme has recently established a new project focusing on Climate and the Cryosphere (CliC) that seeks to understand the role of the cryosphere in the climate system.

INTRODUCTION

The proximity of the Earth's mean effective temperature (255K) to the triple point of water (273K) ensures that water is present in the Earth system in all its three phases—vapor, liquid, and ice. Changes between each phase involve substantial input/release of energy; the latent heat of fusion, required for melting ice, is only 334×10^3 J/kg, while that for vaporization (evaporation) is 2.5×10^6 J/kg. Accordingly, sublimation of snow/ice has a somewhat greater energy requirement than melting. Nevertheless, sublimation occurs when there is a vapor pressure gradient between a snow or ice surface and the atmosphere.

The geological record, spanning nearly one billion years, shows that the Earth has probably experienced the full suite of possible ice regimes. These are summarized in Figure 1. They range from a Snowball Earth (Hoffman et al., 1998; Hoffman and Schrag, 2000) in the Neo-Proterozoic epoch (about 800–600 Ma) to an ice-free state in

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<table>
<tr><td>Snowball Earth</td><td colspan="3">Bi-Polar Ice Age</td><td>Uni-Polar (Antarctic) Ice Age</td><td>Ice Free Earth</td></tr>
<tr><td rowspan="4">Late Proterozoic Glaciations ~800 to 550 Ma</td><td colspan="2">Extensive Ice</td><td>Limited Ice</td><td rowspan="4">Eocene-Pliocene (~20 - 4 Ma)</td><td rowspan="4">Cretaceous Period (~100 Ma)</td></tr>
<tr><td>Polar-Tropical Ice sheets</td><td>Polar Ice Sheets</td><td rowspan="2">Interglacial Periods: Last Interglacial (~125,000 ka) Present Interglacial (10,000 ka - present)</td></tr>
<tr><td rowspan="2">Ordovician Glaciation (~450 Ma) Permo-Carboniferous Glaciation (~280 Ma)</td><td>Late Cenozoic Glaciation (~2.5 Ma - Present)</td></tr>
<tr><td colspan="2">~ 20 Glacial Cycles (Duration ~100 ka after 0.8 Ma)</td></tr>
</table>

Fig. 1. States of the Earth's climate. Abbreviations: Ma = million years; ka = thousand years.

the Cretaceous epoch (Barron, 1983) and in the early Tertiary period (Huber et al., 1999). There have also been two protracted Ice Ages with extensive land ice on the continents. These Ice Ages affected Gondwanaland in the Southern Hemisphere during the late Ordovician–early Silurian (about 440 Ma), and in the Permo-Carboniferous epoch, and probably extended into low latitudes (Hyde et al., 1999). The latter persisted from 330 to 275 Ma. It is assumed that there was extensive sea ice in the northern polar region. The late Cenozoic glaciations of the Plio-Pleistocene epochs have been bipolar, but a unique situation seems to have existed in the Miocene (20–8 Ma), according to Flohn (1987), when there was a major ice sheet in the Antarctic (Sugden et al., 1995), but the North Polar Region was relatively temperate. The extent and nature of the ice cover of Antarctica during the Pliocene remains controversial (Miller and Mabin, 1998). Glaciation is first reported around 7 Ma in Greenland (Larsen et al., 1994) and between 5.0 and 6.7 Ma in southeast Alaska (Lagoe et al., 1993). The latter source indicates warm middle Miocene conditions in the North Pacific and a relatively warm mid-Pliocene interval in Alaska, following the initial glacial phase, until at least 3.5 Ma. The relative roles of forcing external and internal to the climate system in creating these conditions are only partially understood. During the last 800,000 years, there have been recurrent glacial/intergacial cycles lasting about 100,000 years regulated by orbital variations of the Earth. However, bipolar ice has persisted through the short (~10,000 year) interglacial intervals.

Snow and ice cover influences the atmosphere and ocean, and therefore the climate system, through both direct and indirect effects. The most fundamental effect is attributable to the high reflectivity of snow/ice to incident solar radiation (*albedo*) compared with other natural surfaces. Fresh snow has an integrated albedo of 0.8–0.9 and bare ice about 0.6 (Barry, 1996); most land surfaces have an albedo in the range 0.1–0.3, while that for a water surface is as low as 0.05. Climate forcing can result if atmospheric cooling leads to an extension of the snow/ice cover, increasing the surface albedo and thereby leading to more extensive snow/ice, and thus to additional cooling—a positive feedback process. It should be noted that calculations of the albedo effect need to take account of the latitudinal variation of available incoming solar radiation (Robock, 1980). The insolation-weighted effect of areas covered by snow and by sea ice has been determined by Pielke et al. (2000). They show that the albedo effect of snow cover is maximum in April when there is still extensive snow on the northern land areas. For sea ice, which is more persistent, the maximum effect is delayed until May–June when there is more incoming solar radiation. Snow and ice also insulate the underlying land/ocean surface and largely shut off transfers of turbulent energy to/from the atmosphere. For sea ice, most transfers are through leads and other open water areas in the ice.

The specific effects of various components of the cryosphere on the climate system will now be discussed. Three time scales are considered—the geological, centennial-millennial, and interannual-decadal scales—because the relative importance of different processes is to some degree dependent on the time scale under consideration.

GEOLOGICAL TIME SCALE

Here we consider hemispheric to global-scale processes operating over intervals of 10^5 to 10^6 years during geologic time.

Ice Sheets

The effect of ice sheets on the atmosphere has primarily been examined through studies using general circulation models (GCMs), although diagnostic analyses of effects of ice-sheet orography on the atmospheric wave structure and synoptic activity have also been performed. The location of the ice sheet and its size are important factors for climate sensitivity to the presence of the ice (Shinn and Barron, 1989; Crowley et al., 1994). Antarctica is located beneath the polar vortex, whereas the Laurentide ice sheet of the Last Glacial Maximum affected the westerly jet stream and the transient eddies that dominate the poleward transports of angular momentum and energy. At present, Antarctica appears to modulate the phase of planetary wave number one and gives rise to wave number three (James, 1988). Ogura and Abe-Ouchi (2001) examine the relative importance of ice albedo–temperature feedback and ice sheet topography on the atmospheric circulation for Antarctica in a general circulation model (GCM) study. They show that atmospheric cooling over Antarctica is principally attributable to the high albedo and not the high elevation. However, the topography modifies the spatial pattern of the cooling, enhancing it over Antarctica and suppressing it over the Southern Ocean through the reduced poleward heat transport caused by the presence of the high ice sheet. The findings support the concept that the existence of an ice sheet tends to make it self-sustaining.

Experiments for Carboniferous boundary conditions, with a coupled ice sheet–energy balance model (Hyde et al., 1999) reproduce the extensive glaciation of Gondwanaland (the Pangaean mega-continent) with an ice volume about $\times$ 4 that inferred for the Last Glacial Maximum. Earlier experiments with an atmsopheric general circulation model, coupled to a mixed-layer ocean with a sea ice cycle, demonstrated that snow cover on the Pangaean mega-continent would have generated extreme seasonality and mega-monsoon circulations (Kutzbach and Gallimore, 1989).

Numerous GCM experiments for the Last Glacial Maximum (21 ka) have addressed the influence of the Laurentide and Fennoscandinavian ice sheets on the circulation

(Manabe and Broccoli, 1985; Felzer et al., 1996; Bush and Philander, 1999; Kageyama and Valdes, 2001; Roe and Lindzen, 2001). In general, they suggest an increased tendency toward stationary wave/blocking type patterns, and displacement of the storm tracks. The jet stream may be split around the North American ice sheet. Also, there are enhanced tropical wind systems with effects on the ocean temperature structure.

There have been various studies of the climatic factors leading to ice sheet initiation. It remains unclear, however, where the initial snowfields would have formed (Williams, 1978; Koerner, 1980). The time required for significant ice masses to build up can be estimated from the record of sea level changes and is of the order of 5,000–10,000 years. It is also unknown what was the extent of perennial and seasonal snow cover during even the Quaternary glaciations.

Frozen Ground

Permafrost probably developed in high northern latitudes during the Pliocene Epoch (~3.5 Ma). During glacial intervals, when the sea level was lowered by at least 130 m, frozen ground also developed on the exposed continental shelves of the shallow seas off northern Siberia (Rosenbaum and Shpolyanskaya, 2000). Despite postglacial marine transgressions, sub-sea permafrost persists in many areas as a result of low bottom-water temperatures and other factors (Zhigarev, 1997). The primary climatic role of permafrost results from its effects on drainage and the moisture balance. The ground thaws in summer only to a depth of around 20–150 cm, depending on the duration and warmth of the snow-free season and the insulating effect of the winter snow pack thickness and the vegetation cover. Hence, following snowmelt the ground becomes waterlogged, except on steep slopes, and nearly all of the available energy goes into evaporation. This helps suppress air temperatures in tundra regions. The impeded drainage also modifies the annual hydrograph in northern rivers and delays the delivery of fresh water into the Arctic Ocean.

Arctic Sea Ice

New results from seabed mapping by submarine in 1999 strongly support the notion of ice shelves in the Pleistocene Arctic (Polyak et al., 2001), perhaps resolving a longstanding controversy over the presence of ice shelves versus perennial sea ice. Glacial molding and scouring of bedforms, and the fluting or erosion of submarine ridge crests, was found to 1 km depth on the Chukchi Plateau, Northwind Escarpment, Lomonosov Ridge, and Yermak Plateau. Ice shelves would have eliminated any ocean-atmosphere energy exchanges. Their extent and date of their breakup are unknown.

Sea ice may have formed seasonally in the Arctic Ocean during the Pliocene and become perennial at least 750 ka according to Herman (1983). Clark (1982; Gilbert and Clark, 1983) suggests that perennial ice was present before 2 Ma and after 0.7 Ma, with an intervening period of more open/thinner ice conditions. Sea ice plays a major role in winter climate by isolating the atmosphere from the underlying ocean (Barry, 1989). In summer, there is a strong albedo effect. At present, the surface albedo in the central Arctic in May is around 0.80, decreasing to about 0.50 by early August as the snow melts and puddles form on the sea ice (Robinson et al., 1992).

Variability in the timing of snow melt onset can cause an interannual variation in absorbed solar energy estimated to be sufficient to melt 66 cm of ice, compared with an annual ice growth of 2 m by thermodynamic processes (Barry et al., 1989).

Sea ice extent has an important effect on cyclone tracks and areas of cyclogenesis, as shown by model studies where a major polynya is introduced (Glowienka-Hense and Hense, 1992), or the ice cover is removed (Royer et al., 1990). They also find that removing winter sea ice leads to increased precipitation in northern high latitudes. During the Last Glacial Maximum, with sea ice in the North Atlantic reaching about 50°N, mid-latitude cyclones moved more zonally instead of into the Norwegian Sea, and the tracks in both the North Atlantic and North Pacific are shifted eastward (Kageyama et al., 1999).

MILLENNIAL-CENTENNIAL SCALE VARIATIONS

Snowline

Here we are mainly concerned with fluctuations of mountain glaciers and snowline. Reconstruction of glacier fluctuations by Röthlisberger (1985) suggests climatic oscillations, of uncertain cause, on a time scale of about 1000–1500 years. The changes in area of land ice cover were in general rather modest, such that minor albedo effects could be anticipated. Nevertheless, for the Alps of eastern Switzerland, Maisch (1987) calculated a 70 m (±51 m) rise in snow line between 1850 and the 1980s. Hence, the extent of snow cover may have shown more significant variations of climatic importance. The lowering of snow line during the Little Ice Age (AD 1550–1850) interval of glacier expansion is estimated to have been around 100–200 m (Porter, 1986; Grove, 1988). In north-central Baffin Island, increased snow cover extent on plateau areas during a Little Ice Age phase ca. 300–100 years ago is identifiable by lichen-free rock surfaces (Ives, 1962; Locke and Locke, 1977).

Ice Rafting

A process occurring on the (multi-) millennial time scale has been identified from ocean sediments in mid-latitudes of the North Atlantic. Heinrich (1988) first recognized layers of ice-rafted detritus (IRD) dropped by decaying icebergs from massive discharge events in eastern Canada during the last glacial cycle. Bond et al. (1992) showed that between 70 and 14 ka these events occurred at intervals of between 5000 and 10,000 years; the current view is that they occurred about every 6000 to 7000 years. The transport of detritus by icebergs is poorly understood according to Andrews (2000). Moreover, the extent and volume of the detritus implies spatially extensive and prolonged calving events. Clark et al. (2001) propose that when the margin of the Laurentide ice sheet was between about 43° and 49° N, ice-margin fluctuations triggered re-routing of the Mississippi River discharge into the western North Atlantic via the Hudson or St. Lawrence rivers. The freshwater pulses coincide with observed reductions in North Atlantic deepwater formation and thus they affected the thermohaline circulation.

Subsequently, Bond and Lotti (1995) reported more frequent, smaller IRD events between 40 ka and 16 ka that correlated with warm-cold Dansgaard-Oeschger (D-O) cycles in the Greenland ice cores. Later, Bond et al. (1997) showed that these also extend through the Holocene and have a period of 1470 ± 500 years. Schultz et al. (1999) indicate that the amplitude of the 1470-yr oscillations was large when the ice volume gave a 45 m lowering of sea level and there were important ice mass changes. Small-amplitude variations occurred during intervals of relatively stable sea level with reduced climatic variability in the entire 500–5000 yr range, as evidenced by the $\delta^{18}O$ in the Greenland ice cores. The finding of δO oscillations during the Holocene implies some large-scale climatic mechanism, possibly one involving the coupled ocean-atmosphere system. According to van Krefeld et al. (2000), this may have involved meltwater injections in the East Greenland Current, resulting from iceberg discharges associated with internal dynamical processes in the Greenland ice sheet. A potential climatic role of such outbreaks involves the freshwater layer in the western North Atlantic. At present, ocean convection and deep water formation in the Northern Hemisphere takes place in the Greenland Sea and Labrador Sea through mesoscale ocean processes (Carsey and Roach, 1994). Enhanced ice transport via Fram Strait under modern conditions, or from instability of the Laurentide or Greenland ice sheets generating many icebergs during glacial times, could shut off such convection, thus changing the circulation of the global ocean conveyor belt and global climate.

An alternative hypothesis is advanced by Alley et al. (2001). They argue that outburst floods or iceberg discharges are unlikely to have been periodic. However, they suggest that weak periodic forcing within the North Atlantic climate system, augmented by “noise” from ice sheet–related processes could set up stochastic resonance on a 1500 yr time scale. Warming events are predicted at times T, 2T, and 3T, with exponentially decreasing occurrences as the waiting time increases. A histogram of the recurrence of waiting times derived for the Greenland $\delta^{18}O$ record fits the hypothesis. They propose that the Bond cycles, comprising successively colder D-O oscillations, culminate in a Heinrich event.

INTERANNUAL-DECADAL SCALE VARIATIONS

Important variations occur on interannual to decadal scales in the seasonal components of the cryosphere, involving snow cover, seasonal sea ice and seasonally frozen ground. These are the most spatially extensive elements of the cryosphere. Together, seasonal snow and ice cover in the Northern Hemisphere occupies an area of about 58 million km^2 at maximum and reaches a mean latitude of 50°N in January. Mokhov (1984) estimates that this boundary has a temperature sensitivity of about 2 ± 0.5 deg. latitude/K, based on empirical data and model experiments. In part, this must represent an albedo feedback effect, as discussed below.

Snow Cover

At its mean annual maximum extent in February (based on 1972–1997), snow covers an area of 43 million km^2 in the Northern Hemisphere. Fluctuations in extent of the order of 10–15 percent occur both interannually and on a decadal scale (Robinson et al., 1993). There has also been a decrease in the area, affecting primarily Eurasia, especially in spring and summer, from the late 1970s to the early 1990s. This appears to have had a feedback effect on air temperatures, via the absorbed solar radiation, accounting for a warming in northern mid-latitudes of perhaps 0.3°C (Groisman et al., 1994).

The effects of snow cover on the atmosphere and climate have received considerable attention (for example, Walsh, 1987; Groisman and Davies, 2000; Cohen and Entekhabi, 2001). They include a reduction in air temperatures over snow-covered ground and an associated decrease in atmospheric 1000–500 mb thickness (Lamb, 1955). The latter tends to influence the steering of cyclones, which in turn can modify the occurrence of snowfalls (Williams, 1978). Snow cover also insulates the underlying soil and alters the turbulent transfers of sensible and latent heat between the surface and atmosphere. The characteristics of the insulation effect depend on the timing of snow cover formation, its persistence, and thickness.

On a continental scale, there are well-known feedback effects between Asian snow cover extent in spring and the rainfall over India during the summer monsoon, as confirmed by the recent analysis of Kripilani and Kulkarni (1999). The lag effect results from the weakened land-sea temperature gradient set up by the soil moisture deriving from the snow pack. Eurasian snow extent also modifies the downstream planetary wave structure and climate over the North Pacific and western North America, (Clark and Serreze, 2000). Extensive (reduced) snow cover leads to an enhanced (weakened) downstream ridge over the western cordillera of North America with contrasting patterns of climatic anomalies in the two cases.

Sea Ice

Sea ice cover insulates the ocean from the atmosphere, but significant heat fluxes can occur via linear cracks (leads) that open in the ice through divergence. They range in width from tens of meters to more than a kilometer and can refreeze within 24 hours. However, they transfer both sensible and latent heat into the atmosphere in plumes that have been shown to penetrate the arctic inversion (Schnell et al., 1989). Such plumes may account in part for the concentrations of ice crystals (*diamond dust*) observed in the Arctic lower troposphere in winter (Curry et al., 1990.).

The most significant effect of sea ice on climate involves ice-albedo feedback, which is complicated by the snow cover and its summer melt, melt ponds which cover about 25 percent of the central Arctic sea ice in mid-July, and the lead fraction (Curry et al., 1996). On a seasonal time scale, analysis of anomalies in sea ice extent and in atmospheric circulation show that during August–January the ice anomalies and atmospheric anomalies are largely synchronous, whereas during the ice-decay season February–July, atmospheric anomalies precede a response in the ice (Walsh and Johnson, 1979).

Frozen Ground

Regions where permafrost is continuous, discontinuous, or sporadic together represent 24 percent of the exposed land surface of the Northern Hemisphere. Considering the fractional coverage in these zones, permafrost may actually underlie between 13 and 18 percent of the land area (Zhang et al., 2000). In addition, a further 25 percent of the soil surface outside the permafrost zone is seasonally frozen for at least two weeks per year (Barry and Zhang, 2001). This climatological estimate is based on a simple relationship between mean monthly air temperature and the number of days when the ground is frozen. However, operational mapping of surface soil state, when the ground is snow free, also appears feasible using passive microwave remote sensing, calibrated via the air temperature algorithm.

The frozen state inhibits evaporative loss of moisture from the soil, and hence any change in the duration of this condition will affect the atmospheric moisture regime. Such processes are only just being incorporated in land surface parameterization schemes as a prelude to their inclusion in general circulation models.

CONCLUDING REMARKS

The important role played by snow and ice cover in climate is apparent on all time and space scales. The predominant influence is attributable to albedo-temperature feedback, but there are significant other roles. These include the following: ice sheets exert a topographic influence on the atmosphere modifying the planetary wave structure and the distribution of synoptic systems; snow cover and sea ice both insulate the underlying surface and modify the turbulent transfer of heat and moisture into the atmosphere; frozen ground has a similar effect on turbulent transfers; snow cover and permafrost both modify the runoff regime; sea ice export from the Arctic influences the structure of the northwestern North Atlantic and has an effect of the thermohaline circulation.

Because many of these processes are not well determined quantitatively and generally are not well represented, or may even be neglected, in climate models, a new World Climate Research Programme activity to address these issues has been initiated. A science and co-ordination plan for this Climate and Cryosphere (CliC) project has been developed (Allison et al., 2001) and an implementation plan is now being prepared. It is hoped that countries where the cryosphere is a major element in the climate will undertake national programs designed to address questions of special concern. Three phases of activity can be envisaged, relating to the perceived urgency of the questions. In the near term, attention needs to be given to the accelerating loss of tropical glaciers, to ensure that appropriate records documenting their history are collected. In the Andes and even in mid-latitudes, glacier recession has major implications for water resources. On a somewhat longer time scale, model projections of a seasonally ice-free Arctic Ocean within about 50 years, as a result of global warming, need to be carefully assessed. Next, changes in the mass balance of ice sheets and glaciers and their consequent contributions to the global rise in sea level in the 21st century needs to be firmly established. In addition, CliC will seek to foster monitoring of changes in cryospheric elements as indicators of change in the climate system.

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