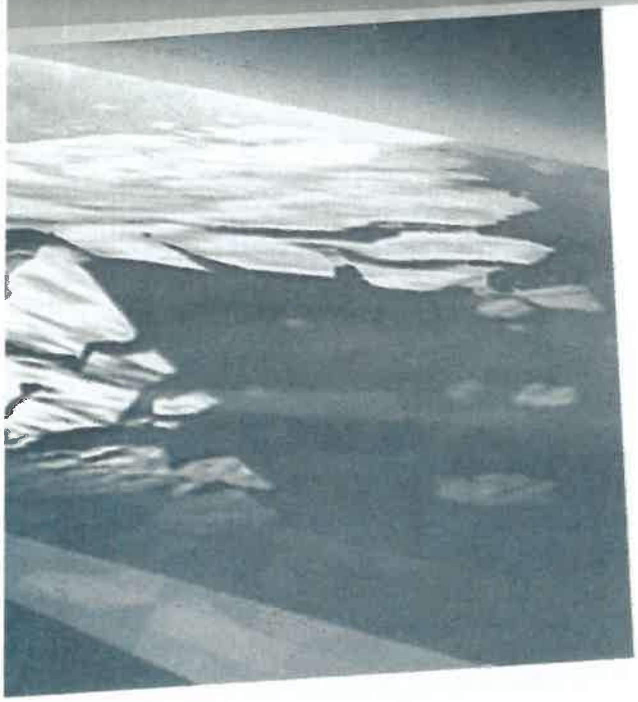


CHAPTER 6

The Cryosphere

Key Questions

- What controls the distribution of sea ice over the oceans?
- How does a glacier move and why is it important?
- What role does the cryosphere play in the global climate system?

Chapter Overview

From the discussion of the global energy balance in Chapter 3 it should be clear that one of the reasons we are interested in the cryosphere is because of its importance in the climate system and, in particular, its very high albedo. Less obvious (unless you live there) is, for example, the challenge that frozen ground presents to living at high latitudes, or the importance of glaciers and mountain snow cover as a source of freshwater. In this chapter, however, we are primarily interested in Earth's major ice caps—Greenland and Antarctica—and in the ice cover of the Arctic and Antarctic oceans. We will talk about how ice forms in both environments (land and ocean), what controls ice dynamics and why it is important, and how both sea ice and the continental ice sheets interact with the climate system over different time scales.

INTRODUCTION

Looking down on Earth from space, the most obvious features of the planet are the oceans and clouds—but equally striking are the large ice masses that occupy both polar regions. Between them, the oceans, clouds, and the polar ice caps (all various forms of water) dominate the global energy balance and play a major role in shaping Earth's climate. The primary components of the cryosphere are the **continental ice sheets** and ice shelves, mountain glaciers, **sea ice**, river and lake ice, snow cover, and **permafrost** (frozen ground). These are illustrated in Figure 6-1 and Table 6-1. We include frozen ground in our definition of the cryosphere but, for the most part, when we talk about the cryosphere in this chapter we are primarily interested in various forms of frozen water at the planet's surface. (Ice particles in clouds are not usually considered as part of the cryosphere.)

The cryosphere interacts with the climate system in a variety of ways and over a wide range of time scales. Changes in the distribution of sea ice and snow cover change the albedo and feed back to regional and global temperatures. Changes in the amount of glacier ice affect global sea level, the melting of permafrost releases greenhouse gases to the atmosphere, and the process of sea-ice formation helps increase the salinity of the surface ocean at high latitudes. This, in turn, affects ocean density, bottom-water formation, and the deep-ocean circulation. Beyond these global-scale processes, sea ice and permafrost play an important role in regional ecosystems, and mountain snow cover and glaciers are an important source of freshwater. As an

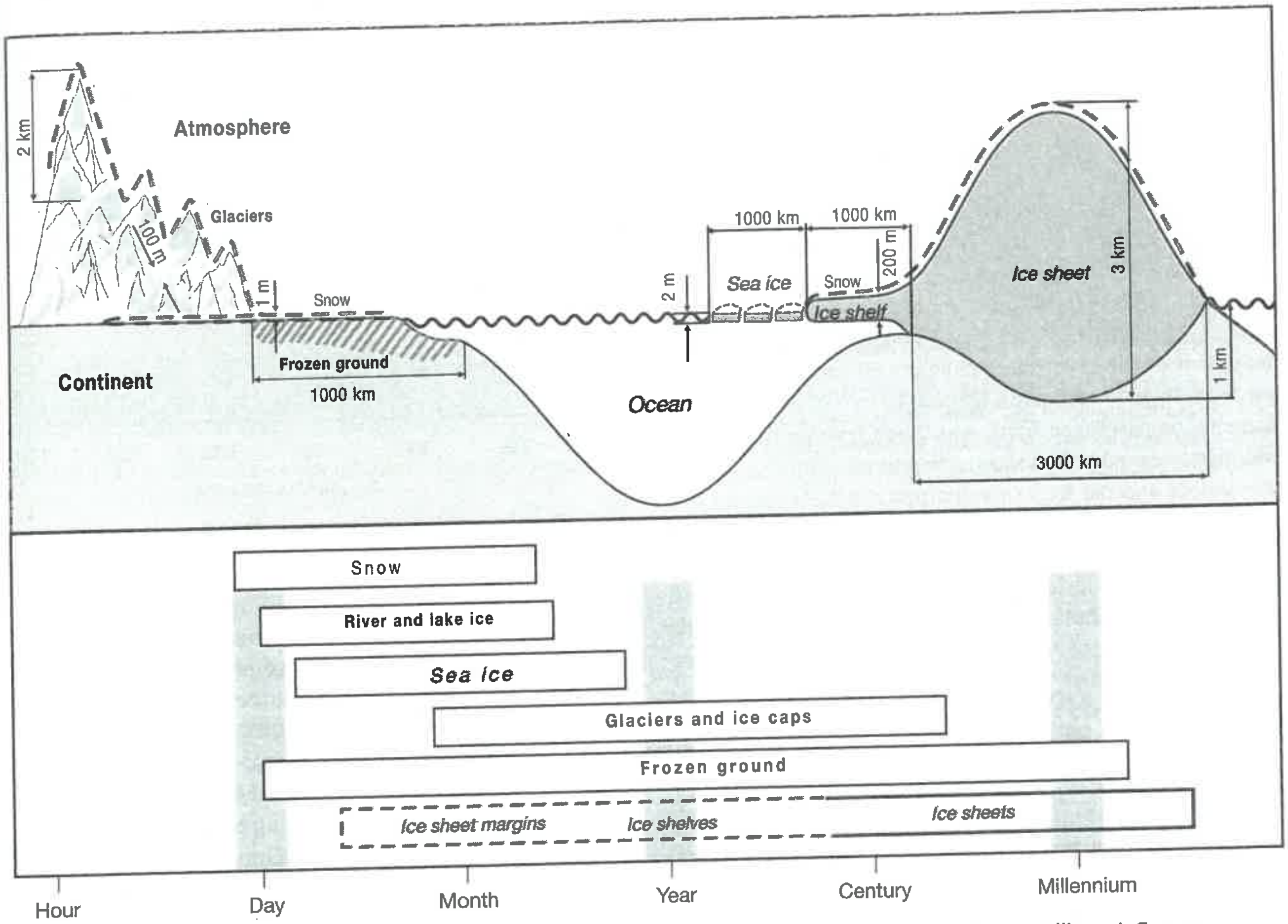


FIGURE 6-1 Components of the cryosphere and their time scales. (*Source:* Lemke, P., J. Ren, R. B. Alley, I. Allison, J. Carrasco, G. Flato, Y. Fujii, G. Kaser, P. Mote, R. H. Thomas, and T. Zhang, 2007: "Observations: Changes in Snow, Ice and Frozen Ground." In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, M. Tignor, and H. L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom, and New York, NY, USA.)

TABLE 6-1 Area, Volume, and Sea Level Equivalent (SLE) of the Cryosphere

Cryosphere Component	Area (10^6 km^2)	Ice Volume (10^6 km^3)	Potential Sea-Level Rise (SLE) (m)
Snow on land (NH*)	1.9–45.2	0.0005–0.005	0.001–0.01
Sea ice	19–27	0.019–0.025	~0
Glaciers and small ice caps			
Smallest estimate	0.51	0.05	0.15
Largest estimate	0.54	0.13	0.37
Ice shelves	1.5	0.7	~0
Ice sheets	14.0	27.6	63.9
Greenland	1.7	2.9	7.3
Antarctica	12.3	24.7	56.6
Seasonally frozen ground (NH)	5.9–48.1	0.006–0.065	~0
Permafrost (NH)	22.8	0.011–0.037	0.03–0.10

*Northern Hemisphere

Source: Lemke, P., J. Ren, R. B. Alley, I. Allison, J. Carrasco, G. Flato, Y. Fujii, G. Kaser, P. Mote, R. H. Thomas, and T. Zhang, 2007: "Observations: Changes in Snow, Ice and Frozen Ground." In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Averyt, M. Tignor, and H. L. Miller [eds.]). Cambridge University Press, Cambridge, United Kingdom, and New York, NY, USA.

example, the mountain snowpack contributes about 75% of the annual surface water supply for the western United States. This moisture might still fall as rain, of course, if temperatures were warmer. However, it would not be available year-round unless it was trapped in some system of reservoirs. The mountain snowpack acts as a large regional reservoir, and this makes it extremely important to the American West and to many other parts of the world.

Like the atmosphere and oceans, the cryosphere is a dynamic system. The major components of the cryosphere (glaciers/ice sheets and sea ice) are in constant motion, and the extent of the cryosphere changes on time scales ranging from days to millennia (Figure 6-1). There have been times in the past when Earth was much warmer than today and there were no polar ice caps, and times when the system was colder and the ice expanded much further toward the equator. There were even times when most, if not all, of the planet was frozen at the surface. These more extreme cases will be discussed in later chapters. Here, we focus on the present-day distribution of snow, ice, and frozen ground.

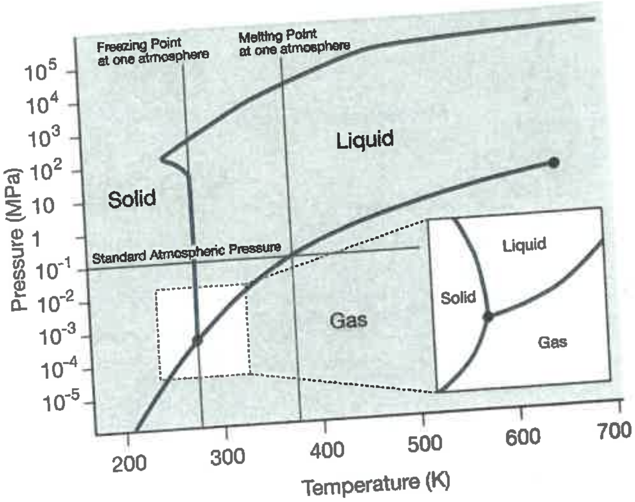


FIGURE 6-2 Phase diagram for water.

RIVER AND LAKE ICE, SEASONAL SNOW COVER, AND PERMAFROST

The seasonal freezing and thawing of lakes and rivers plays an important role in local ecosystems and impacts a range of human activities, but plays only a minor role in the global-scale processes that are the primary object of this text. Other than to point out that they are still part of the cryosphere, we will not discuss them further. For the most part, this chapter will focus on glaciers and ice sheets and on sea ice. However, we will make brief mention of snow cover because of its interaction with the atmosphere and its possible effects on interannual climate variability, and we will discuss permafrost because of its significance as a potential source of greenhouse gas (methane) emissions.

Formation of Snow

Snow forms initially as ice crystals in clouds. In the same way that water droplets form around cloud condensation nuclei (Chapter 4), small ice crystals also form around tiny mineral or organic particles in the atmosphere, referred to as **freezing nuclei.** At that point, the ice crystal can grow and take a variety of forms, which are largely a function of temperature. The dominant process is one of deposition: the ice forms directly from water vapor, rather than from the freezing of liquid water. In Chapter 4 we discussed the different phases of water—gas (water vapor), liquid, and solid (ice)—and we also showed that water could transition directly from ice to vapor (sublimation) or from vapor to ice (deposition) without going through a liquid phase in between. The conditions under which this is possible are shown in the **phase diagram** for water in Figure 6-2. The diagram shows the way water behaves at different combinations of temperature and pressure. The solid lines are the phase boundaries and show the combinations of temperature and pressure at which the two phases can coexist in a stable state. Along the gas–liquid boundary, for example, water can exist as 100% gas or 100% liquid or any combination in between. The horizontal gray line shows the standard atmospheric pressure and the two vertical gray lines show the freezing and melting points of water at that pressure (273 K and 373K, or 0°C and 100°C).

The insert shows an enlargement of the region where all three phase boundaries come together. The curves in this case are not drawn to scale and the slopes are exaggerated to make the behavior more obvious. From the insert you can see that as pressure decreases, the melting point also decreases. Similarly, you can see that at one atmosphere (the standard pressure at sea level), if the pressure is reduced, as happens when an air parcel rises in the atmosphere, then the boiling point is also lowered. There are several other interesting features in these curves. The point at which all of the phase boundaries meet is called the **triple point;** here, water can exist in a stable state in all three phases. The point at which the gas–liquid boundary stops is called the **critical point.** At temperatures and pressures beyond this point there is no real distinction between liquid and gas. The diagram also shows that at high enough pressures ice will not melt, even at extremely high temperatures.

For our discussion of snow crystals, the relevant part of the diagram is the phase boundary between gas and solid. Once the ice crystal forms, it continues to grow primarily by deposition. In the absence of freezing nuclei, pure water will not freeze until temperatures in the cloud drop to 233 K (−40°C). This means that once temperatures drop below 273 K and some ice starts to form around freezing nuclei, the cloud will consist of a mixture of ice crystals and **supercooled water** droplets. (Liquid water that is colder than 273 K is referred to as supercooled water.) The saturation vapor pressure for ice and liquid

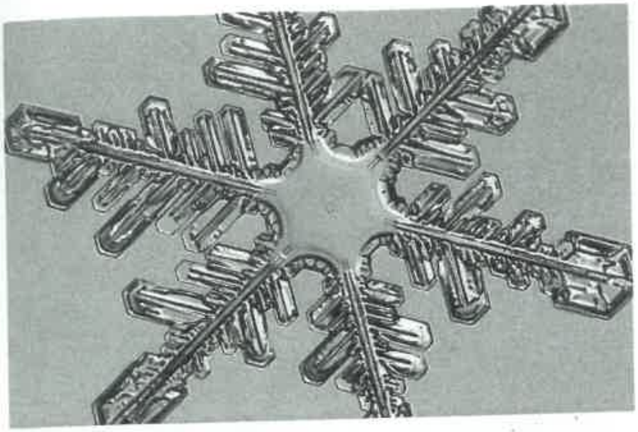


FIGURE 6-3 Characteristic hexagonal structure of a snow crystal. (*Source:* Kenneth G. Libbrecht.)

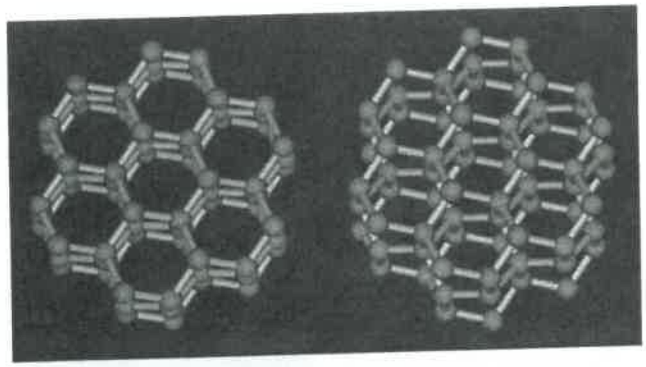


FIGURE 6-4 The hexagonal structure of an ice crystal. The balls represent oxygen atoms while the gray sticks represent hydrogen atoms (the two structures show different views of the same crystal). (*Source:* Kenneth G. Libbrecht.)

water are the same at 273K. As the temperature drops, however, the saturation vapor pressure with respect to ice decreases. This means that as far as the ice is concerned, the air is supersaturated and water vapor will deposit directly on the ice to grow the ice crystal. As water vapor is deposited, the vapor pressure of the air drops, which results in evaporation from the water droplets—and so the ice crystals grow at the expense of the water droplets. In other words, the ice grows by direct deposition and moves from gas to solid without going through the liquid phase first. It is this process that gives rise to the commonly observed hexagonal snowflakes such as those in Figure 6-3. If the ice crystal comes into contact with supercooled water droplets, the water can freeze directly to the ice. In that case, instead of the intricate hexagonal crystal structure, one simply gets rounded ice grains (known as graupel). There is little cohesion between these grains and on the surface; if they are covered by further layers of snow, they can form an unstable layer that can lead to avalanches on slopes. A third way the ice can grow is by agglomeration, as ice crystals join together to form larger snowflakes.

Ice can take on a variety of crystalline structures, depending on the conditions of formation. The most common form at normal atmospheric temperatures and pressures is the hexagonal structure shown in Figure 6-4, which shows two different views of the same crystal. In this diagram, the spheres represent oxygen atoms and the gray connections represent hydrogen atoms. It is this crystal structure that gives ice the properties that make snow cover of interest to us in terms of its interactions with the climate system.

Northern Hemisphere Snow Cover

The snow cover in the Northern Hemisphere varies in area from about $2 \times 10^6\,km^2$ to $4 \times 10^6\,km^2$ in summer to about $45 \times 10^6\,km^2$ to $47 \times 10^6\,km^2$ in winter, during which it occupies about half the land area north of 20°N. The most obvious feature of the snow cover is its high albedo (bright white color). Although individual ice crystals are clear, each crystal is a reflecting surface. When solar radiation hits the crystal, some of the radiation is transmitted through the ice and some is reflected from the surface. The large number of reflecting surfaces means that much of the radiation that falls on the surface is scattered within the snowpack and reflected back to space. A fresh, low-density snow cover has an albedo of about 80 to 90%. Most objects preferentially absorb particular wavelengths of radiation—green leaves, for example, absorb most of the visible light that falls on them—and, therefore, have a relatively low albedo. Leaves also contain pigments that preferentially absorb blue and red wavelengths; hence, they reflect more in the green part of the spectrum and appear green in color. By contrast, ice crystals absorb and scatter all wavelengths of visible light almost equally, as do water and ice droplets in clouds, which is what gives them their white color. In fact, ice does absorb slightly more at longer wavelengths (i.e., the red part of the spectrum), so the deeper one goes into the snowpack (or into a glacier), the more red has been removed, and so the snow (ice) takes on a slight blue color.

The high reflectivity of fresh snow means that it can have a significant effect on the regional energy balance. Figure 6-5 shows a mosaic of satellite images collected over a one-week period in February of 2002. The images show extensive snow cover over a large part of North America, and the contrast between the snow-covered and the non-snow-covered areas is very clear. The high albedo means that most of the incident sunlight is reflected and little is absorbed to heat the snowpack. This allows the snow to persist for long periods if the air temperature remains below freezing. The fact that the snow has a much higher albedo than the underlying surface also means that radiation that would otherwise warm the surface is now being reflected away, making regional temperatures lower than they would otherwise be. The snow boundary in Figure 6-5

FIGURE 6-5 North American snow cover, February 2–9, 2002. A mosaic of NASA images from the Moderate Resolution Imaging Spectroradiometer (MODIS), flying aboard NASA's *Terra* and *Aqua* satellites. (*Source:* NASA.)

extends southwestward from the Great Lakes toward New Mexico. As most of the airflow over the central and eastern parts of the United States comes from the west and northwest, much of it is coming out of this snow-covered region and, therefore, tends to be colder than it would be without the snow cover present. An extensive continental snow cover such as we find over North America or Siberia during wintertime also builds some **thermal inertia** into the climate system. The high albedo reflects energy away that would otherwise be absorbed. But recall from Chapter 4 that it takes 335 kJ of energy to convert a kilogram of ice into water. That is energy that would otherwise be warming the surface, so again it slows the temperature increase from winter to spring.

While the snow cover keeps air temperatures lower, it has the opposite effect on the underlying surface. Fresh snow is an agglomeration of hexagonal ice crystals that contain large amounts of air spaces between the crystals. This air cannot move very readily through the snowpack and so the snow is not a very efficient conductor of heat—it has low thermal conductivity (see Chapter 4). The result is that the snow cover reduces the heat loss from the ground to the overlying atmosphere. As the snow ages, the ice crystals become more spherical and more closely packed (snow density increases). This has the effect of reducing both the albedo and the insulating properties of the snow, although both are usually still higher than if no snow was present.

Permafrost

Permafrost is permanently frozen ground and is defined simply in terms of temperature. (In other words, ice does not actually have to be present.) Permafrost is considered to be present if the ground remains at or below 0°C for

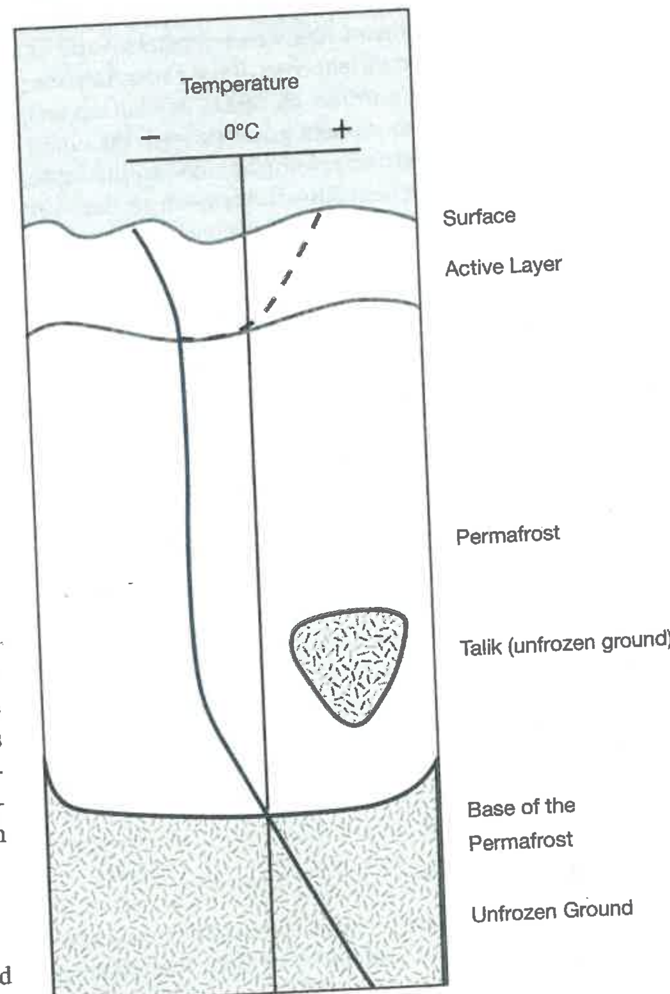


FIGURE 6-6 Sketch cross section of permafrost showing the temperature profile through the different layers.

2 or more years. Figure 6-6 gives a schematic view of what permafrost looks like in cross section. There may be pockets of unfrozen ground (Talik) either at the top of the permafrost or within the permafrost. The permafrost is heated from below and cooled from the surface, and the base of the permafrost is the depth at which the net effect produces a temperature of 0°C. Notice in Figure 6-6 that the temperature at some depth below the surface represents the long-term mean air temperature. Temperature changes in permafrost are slow enough that seasonal and interannual variations in air temperature are averaged out and the temperature is actually a response to the decadal and longer-term temperatures. Lower down, the temperature increases, crossing the freezing point at the base of the permafrost. Near the surface, temperatures may fluctuate widely on a seasonal basis. Surface temperatures are low in winter, but increase in summer. This upper layer, referred to as the **active layer,** freezes in winter and thaws in summer. One can imagine that if there was a long-term cooling trend at the surface, over time the temperature in the permafrost would decrease and the base of the permafrost would move downward. Conversely, if surface temperatures were to increase, then eventually the temperatures within the permafrost would increase and the base of the permafrost would move upward. Measuring the temperature profile in boreholes in the permafrost can, therefore, give an indication of long-term temperature change.

The Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) assembled permafrost borehole temperature data from permafrost regions across the Northern Hemisphere. These data indicate some degree of permafrost warming across most of the region during the second half of the 20th century. Much of this is attributed to the direct effects of increasing air temperatures, while in some cases there also appears to be a contribution from the insulating effects of snow cover. The distribution of permafrost has also changed over time and, in North America, there is evidence that the permafrost is moving northward in response to long-term warming since the Little Ice Age (see Chapter 15). The IPCC Fourth Assessment Report suggests that the permafrost area in the Northern Hemisphere is likely to decrease by 20–35% by the middle of the 21st century. Much of this will be due to changes in the distribution of relatively thin and discontinuous permafrost on the high-latitude margins.

Where the permafrost has high ice content, thawing can cause the ice to melt and the surface to subside. Such pockets of subsidence result in a heterogeneous landscape of depressions and small hills—referred to as **thermokarst.** The higher areas tend to dry out while the hollows accumulate water, which affects local ecosystems, and both long-term changes in topography as well as seasonal freezing and thawing create challenges for construction and infrastructure development. In the context of this book, however, the greatest concern over melting permafrost is the potential for increasing greenhouse gas emissions to the atmosphere.

Near-surface melting of permafrost results in lakes and water-logged soil with anaerobic (low oxygen) conditions that create an environment in which methane-producing organisms can flourish. As we will see in Chapters 12 and 15, methane is an important greenhouse gas, and the atmospheric concentration has more than doubled over preindustrial levels. Chapter 15 also shows that the rate of increase in atmospheric concentration has slowed or stopped in recent years. This is potentially good news for greenhouse warming; however, melting permafrost could change this outlook. Methane emission from regions of melting permafrost could potentially double to over 100 million metric tons per year. However, there is considerable uncertainty in these numbers and some estimates suggest that methane emissions could go much higher. The Arctic is already warming much faster than the rest of the globe due to positive feedbacks such as the snow and ice–albedo feedback introduced in Chapter 3. As permafrost continues to melt, there is the potential for more methane release and a positive feedback that will increase temperatures and further accelerate melting. The potential for this is even greater when one considers methane clathrates that are found in offshore permafrost. Clathrates are solid compounds that can trap a gas molecule within them. In this case, the methane is surrounded by a lattice of water molecules and is referred to as a methane clathrate hydrate. These form at relatively low temperatures and high pressures and are found extensively on the continental shelves, not just in polar regions. They represent a large potential source of greenhouse gases (see Chapter 15). Increased atmospheric methane due to permafrost melt is not currently included in the global climate models used in the IPCC climate change assessments discussed in Chapter 16.

GLACIERS AND ICE SHEETS

If snow cover persists through the summer and starts to accumulate over time, the snow increases in thickness, but also undergoes various transformations. Temperature gradients in the snowpack, together with differences in the size and shape of the ice crystals (e.g., convex or concave surfaces), cause localized differences in vapor pressure that result in sublimation from some crystals and deposition on others. This tends to produce rounding of the crystals and compaction, which increases density. The compaction is further increased by pressure as the snowpack increases in thickness. The ice crystals fuse together where they contact each other and they bond through a process referred to as pressure sintering. As the snow is compacted and the ice crystals fuse together, the density increases, the volume of air between the ice grains is reduced, and the snow is transformed into glacier ice. In cold glaciers where temperatures never come close to the melting point, this process can take hundreds to thousands of years (particularly in regions—such as central Antarctica—where snow accumulation may be only centimeters per year). Where the surface is subjected to melting, however, the meltwater can percolate down through the pack and refreeze, considerably speeding up the process of glacier ice formation. In these conditions, with high snowfall and frequent melting–freezing cycles, the transformation from snow to glacier ice may take only a few years. New snow may have densities of about 50–70 kg m^{-3}. As it goes through the transformation to glacier ice, in the intermediate step it is referred to as **firn** (which has densities of about 400–800 kg m^{-3}), and once the air cavities are sealed and the ice becomes solid, it becomes glacier ice with densities of about 850–900 kg m^{-3}. At this stage, the ice has low permeability and it can flow and move under its own weight. This low permeability and the ability to flow are characteristics that make glaciers, especially the large continental ice sheets of Greenland and Antarctic, particularly interesting from an Earth systems perspective.

Glacier Flow

Glaciers are broadly categorized into mountain (or alpine) glaciers and continental glaciers. (Other categorizations exist, but they are not relevant to the discussion here.) **Mountain glaciers,** as their name suggests, are found in mountainous regions such as the European Alps, the Himalayas, and the Andes. These glaciers cover relatively small areas and are largely confined to mountain valleys. Our focus here is primarily on the large continental glaciers such as those that are found today on Greenland and Antarctica. These two continental glaciers cover extensive regions and are not confined by the topography. Between them, they account for about 97% of the surface area covered by land ice and about 99.6% of the ice volume. As one can see from Table 6.1, most of this ice is in Antarctica. If Antarctica were to melt, global sea level would increase by about 56 m. Melting all of Greenland, by comparison, would raise sea levels by about 7 m. One thing all of these glaciers have in common, however, is that they are big enough to move.

Most of us are familiar with small blocks of ice—ice cubes that you put in drinks. That ice is brittle; when it is stressed (if you hit it with something hard) it will fracture. At higher pressures, however, deeper within a glacier, slowly applied stress causes the ice to deform more like a plastic. This stress increases as you move deeper into the ice and causes the ice to flow (see the Box "Thinking Quantitatively: Movement of Glaciers"). If the ice is frozen to the bed, then the flow at the base is zero and the maximum flow is somewhere above the influence of the bed. Similarly, friction at the sides in a valley glacier (a glacier flowing down a mountain valley) reduces the flow at the edges, and so the maximum flow is toward the center of the glacier and between the surface and glacier bed (Figures 6-7a and 6-7b). Notice in 6-7a that the plastic deformation does not extend up to the surface. Above a

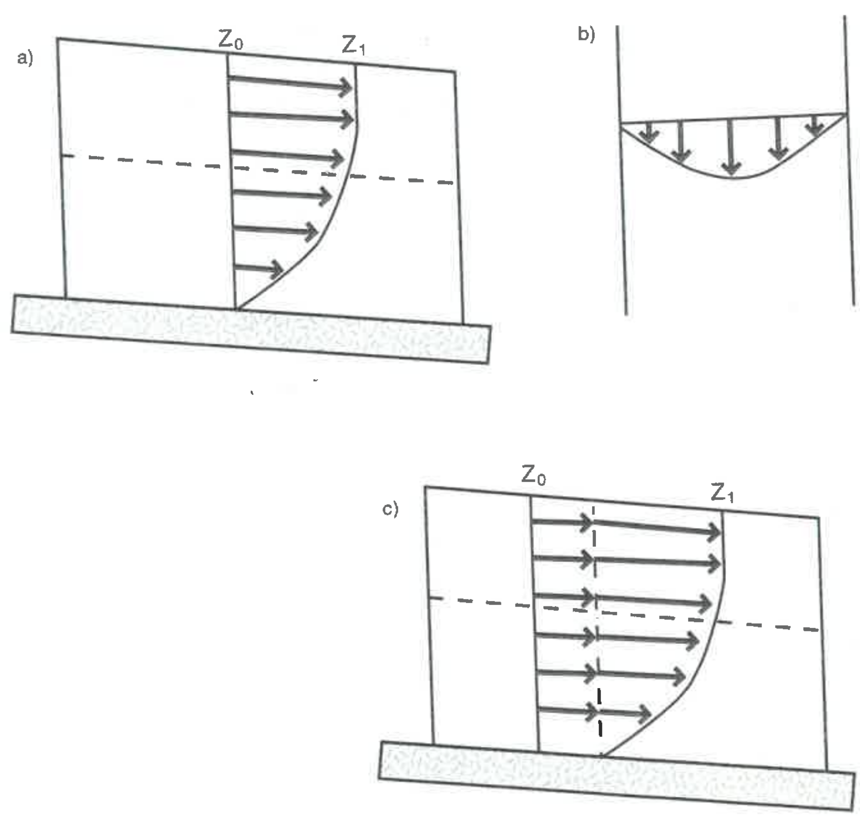


FIGURE 6-7 Ice flow profile: (a) longitudinal section; (b) plane view; and (c) view including both plastic flow and basal sliding.

THINKING QUANTITATIVELY

Movement of Glaciers

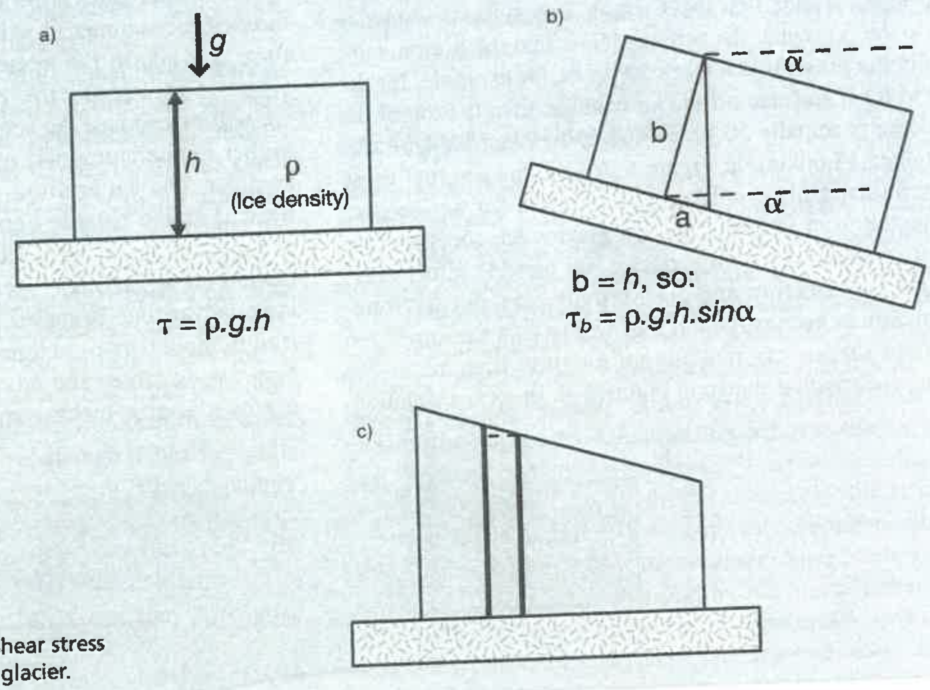
Box Figure 6-1a shows the simplest case for a glacier, in which a large rectangular block of ice is sitting on a flat surface. As one moves deeper into the ice, the pressure from the overlying ice increases. At any point in the block, a **stress** is exerted on the ice by the ice above it. This stress is determined by the weight of ice pushing down on it. Weight is simply the force of gravity g acting on an object of a particular mass. In this case, if we think of a thin column of ice over a unit area ($1 m^2$), then the mass is given by the density of the ice ρ multiplied by the height of the column h. The stress τ is then given by

$$\tau = \rho g h.$$

If we now place the block of ice on a slope (Box Figure 6-1b) we can see that there are two components to the stress: a **normal stress** that acts perpendicular to the slope and a shear stress that is parallel to the slope. The length of lines a and b give the relative magnitudes of these two components of the stress. As these three lines form a right angle triangle, we must have $\sin\alpha = a/b$, and $a = b \sin\alpha$. Since b is the thickness of the ice h, so the shear stress across the base of the column (the **basal shear stress**) is equal to

$$\tau_b = \rho g h \sin\alpha.$$

For small angles, one can show that the effect of the bed slope is negligible, and what is important is the surface slope of the glacier. Looking at Box Figure 6-1c, we see that if we consider a thin column of ice, h is slightly higher on the upslope side, which means that the shear stress (force) is slightly higher on the upslope side than on the downslope side, and that difference causes the ice to deform and flow in the downslope direction (even if the bedrock is level). This deformation or flow occurs by breaking the bonds between layers in the ice crystal lattice. Looking back at the three-dimensional view of the ice crystal in Figure 6-4, we see that the planes are held together by hydrogen bonds. The stress breaks these bonds, causing the ice crystals to slide forward and down, where the bonds reconnect.



BOX FIGURE 6-1
Relationship between basal shear stress t_b and the surface slope of a glacier.

depth of about 50 m, the stress is not large enough to cause plastic deformation. In this case, the ice near the surface is simply being carried along by the ice flow at depth. Where the rate of flow changes, maybe due to a large change in the subsurface topography, then the brittle ice near the surface fractures, forming a crevasse. Where the ice isn't frozen to the bed—where there is liquid water at the base, or where the bed is made up of unconsolidated material (particularly if that also has a high water content)—then the glacier can slide over the bed. In this case, the downslope movement has two components, one due to plastic deformation and one to basal sliding (Figure 6-7c).

So how does glacier movement play into the themes of this book? First, we already know from Chapter 1 that

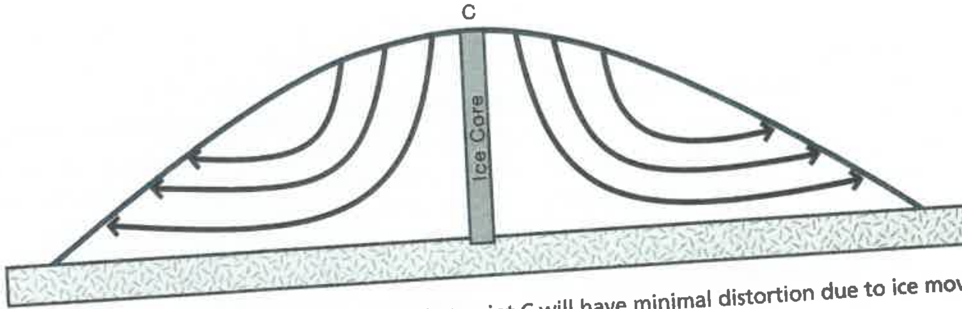


FIGURE 6-8 Flow lines on an ice dome. An ice core drilled at point C will have minimal distortion due to ice movement.

the major ice sheets such as Greenland and Antarctica are an important source of information on past climates and past atmospheric gas concentrations. During the transformation of snow to ice already described, the air spaces present in the snow become sealed forming small air cavities in the ice. When we drill a core down through the ice, we can extract the gas from those cavities and determine the atmospheric gas concentrations at the time this transformation occurred. It should now be obvious, however, that we can't simply drill anywhere on the ice cap. Reconstructing the history is much easier if we know that the ice at the bottom of the core is the oldest, that the ice immediately above that was laid down in the same place and slightly later, and that there is a continuous sequence up to the present at the surface. (Note that the present isn't really the present! If it takes 50 years, for example, for the ice to form and seal off the air cavities, then the top of the ice core is actually 50 years older than when the core was obtained.) Looking at Figure 6-7c, we can see that these criteria certainly would not be satisfied for an ice core extracted from a rapidly moving glacier. Ice cores extracted from anywhere on the glacier can provide information on the ice structure and glacier dynamics, but for long-term climate reconstruction it is best to take the core from parts of the ice cap that are not moving. Imagine an ice dome such as that shown in Figure 6-8. In such a situation, the flow would be radially away from the highest point at the center of the dome. The flow would be fastest where the surface slope is greatest, and minimal where the surface slope is zero—a core drilled at point C would have minimal problems due to deformation and movement. The new deep core in Antarctica that was mentioned in Chapter 1 was drilled at the top of dome C at the dynamical center of the East Antarctic ice sheet.

The movement of glaciers, particularly the big continental ice sheets, has other, more direct impacts on the Earth system. A glacier is not a uniform mass. Snow falls on the glacier in winter, and some of that melts in summer. Where temperatures are cold enough that the snow persists through the summer, it will build up over time and transform into ice in the **accumulation zone** of the glacier (Figure 6-9). Because the ice will then flow downward, the ice will extend past the accumulation zone into regions of higher temperatures where melt (ablation) will occur in summer. The ice profile—its thickness and extent—will depend on the balance between accumulation, transport, and melt. If any of these parameters change, the profile of the glacier will change—getting thicker or thinner, retreating or advancing. Because climate tends to be highly localized in mountainous regions, it is possible for a glacier to be growing larger and advancing in one valley, while a nearby glacier is melting and retreating. With widespread

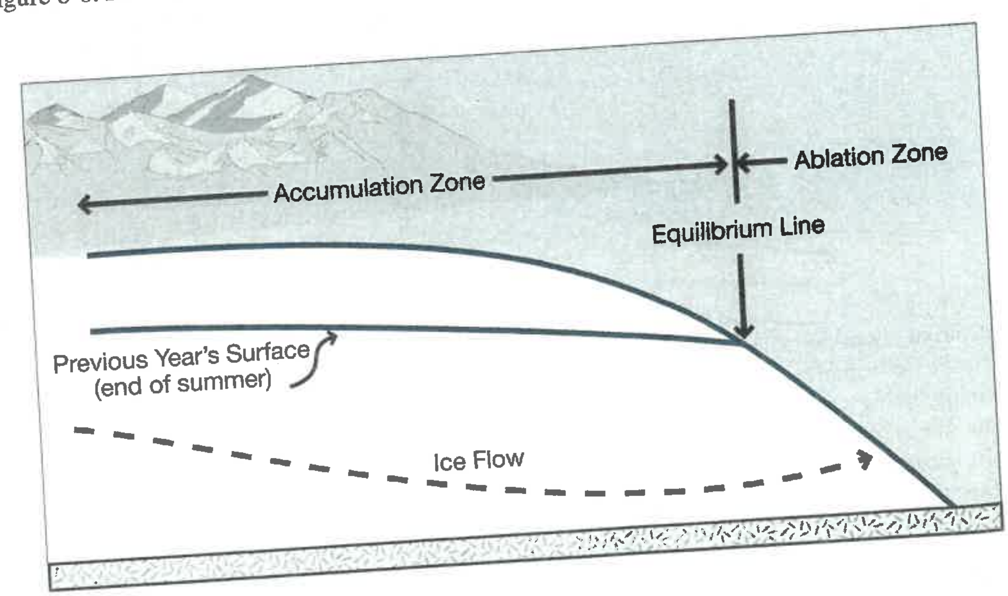


FIGURE 6-9 Accumulation and ablation zones on a glacier.

warming over the latter part of the 20th century, however, most mountain glaciers are retreating.

Imagine what would happen in the polar ice sheets if there was no ice flow. There would be no **ablation zone** because there would be no ice flowing from further inland, and any snow falling in that region would simply melt. Conversely, there would be no melt in the accumulation zone, and any snow that fell would build up as ice. Does this mean the ice in this region would continue to thicken indefinitely? Not really! Once it reached a depth of about 4 km, the ice at the base would start to melt as a consequence of the high pressure, along with warming provided by geothermal heat coming from below. The high pressure at the base of the ice lowers the melting point by a few degrees. But that, by itself, would not make the ice melt, as the temperature at the surface of the ice is about (–50°C in the polar regions. If the ice was this cold at the base, the small freezing point depression caused by the high pressure would not make much difference. What makes the ice melt is the geothermal heat coming from below. This heat flux, which is approximately 0.03 W/m^2 in continental areas (see the next chapter), must get out through the ice by conduction; hence, the temperature must be higher at the bottom of the ice sheet than at the top. When the ice thickness approaches 4 km, the temperature at the base approaches 0°C (or a few degrees lower), and the ice begins to melt.

In reality, ice does move, often aided by basal melting. Under any particular climatic regime, the ice will reach an equilibrium state. Even slight changes in climate, however, can have a significant impact on the ice sheet dynamics. The edge of the Greenland ice cap, for example, forms valley glaciers that end in the ocean. Although the precise mechanism hasn't been fully explained, it appears that higher temperatures during the latter part of the 20th century have caused meltwater to percolate down beneath these outlet glaciers, causing them to flow faster and to increase the rate of ablation and ice loss from the continent. Because of this, Greenland melt is occurring faster than expected. While this isn't yet enough to have a significant impact on sea-level rise or on North Atlantic salinity during this century, we also don't know whether this rate of melt will remain the same, or whether it could significantly increase as warming continues. The potential impacts of a reduction in North Atlantic ocean salinity are discussed further in Chapter 14.

Other surprises could be in store when we look at what could happen to the West Antarctic ice sheet (West Antarctica is the panhandle part of the Antarctic continent). Glacial flow rates are highly variable. They depend on a range of factors such as ice structure, temperature, slope, bedrock type, and the presence or absence of water at the base of the glacier. The velocity of a glacier over its base may vary by a factor of 100 between different glaciers, but typical values range from tens to hundreds of millimeters per day. Some glaciers, however, have been observed to "surge." Surging glaciers move episodically at velocities that may be 100 times faster than normal. If the West Antarctic ice sheet were to surge, it would deliver ice into the ocean at a much higher rate than present, causing a more rapid rise in sea level. While West Antarctica isn't known to surge, the ice sheet is inherently unstable and is held in place now by two large ice shelves at its edges. If the ice shelves were removed (if warming oceans were to melt the ice shelves at their base, for example), then much of the West Antarctic would flow rapidly out to sea (see Chapter 16 for further discussion).

SEA ICE AND CLIMATE

In the last part of this chapter we turn to the polar oceans to discuss the formation of sea ice and to examine climate variability induced by sea-ice effects on atmosphere–ocean interactions at high latitudes. In earlier chapters we hinted at the importance of the ice distribution when we discussed the ice–albedo feedback and bottom-water formation. In this section we will discuss the seasonal distribution of sea ice, and we will examine some of the ways in which changes in sea-ice cover affect the polar ocean climate. These effects are particularly interesting because, as we will see in Chapters 15 and 16, they tend to dominate the high-latitude response to increasing atmospheric CO_2 levels in global climate models.

Ice–Climate Interactions

The seasonal distribution of sea ice over the polar oceans is shown in Figure 6-10. The seasonal range of the ice cover is extreme: The Northern Hemisphere ice cover almost doubles in size from approximately $8.5 \times 10^6\,\text{km}^2$ to $15 \times 10^6\,\text{km}^2$ between summer and winter; the Southern Ocean ice cover grows from $4 \times 10^6\,\text{km}^2$ to $20 \times 10^6\,\text{km}^2$. These are the average distributions measured during the last 30 years of the 20th century using satellite observations of the microwave radiation emitted from the ice surface. Because certain wavelengths of microwave radiation are not attenuated (absorbed or scattered) by the atmosphere and the cloud cover, we can obtain accurate measures of ice extent regardless of the weather. And, because we are measuring radiation emitted by the ice, whether the observations are made during day or night makes no difference. As we indicated in Chapter 1, however, these same data reveal that the summer ice extent in the Northern Hemisphere has been decreasing dramatically during the first decade of the 21st century. If it continues, would we expect this reduced ice cover to have a measurable impact on climate? This is a question we ask in the "Critical-Thinking Problems" section at the end of the chapter. If you read the rest of this section and put it together with what you learned about the atmospheric circulation in Chapter 4, you should be able to suggest some possible answers for yourself!

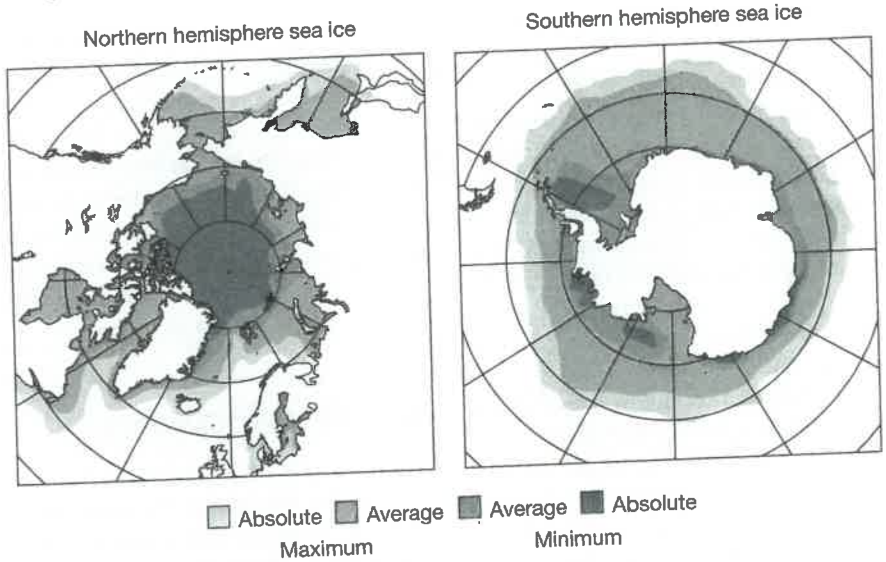


FIGURE 6-10 The seasonal distribution of sea ice in the Northern and Southern hemispheres.

Sea ice forms when the temperature of the ocean surface drops below the freezing point (about −1.8°C for typical ocean salinities). Sea ice grows in thickness as new ice formed from seawater freezes onto the bottom of the icepack. Note that this is very different from how glacier ice forms on land, where the ice forms and accumulates at the surface. New sea ice typically reaches thicknesses of about 1 m during a single winter. The ice cover melts and freezes seasonally at the equatorward margins of the pack but lasts throughout the year at higher latitudes. This permanent pack ice in the Arctic Ocean reaches equilibrium thicknesses of about 5 m; the Southern Ocean ice is much thinner because very little of it survives through the summer melt.

Several features of this ice distribution are important to our discussion. The large seasonal range is a result of thermodynamic controls on the ice cover (that is, the ice cover is responding to the seasonal changes in temperature). The asymmetric nature of the ice margin in the Arctic Ocean, however, reflects the influence of the continental configurations and the ocean circulation: the ice margin extends well to the south in both the western North Atlantic and western North Pacific. At the eastern margins of both oceans, open water extends much farther north off Norway and Alaska. Note that the shape of the ice margin reflects the pattern of ocean currents discussed in Chapter 5. The warm waters of the North Atlantic Drift and the Kuroshio Current prevent the ice margins from extending farther south in the eastern oceans, whereas the East Greenland Current, the Labrador Current, and the Oyashio Current bring colder water and carry increased ice cover southward in the west.

The ice distribution helps illustrate one of the important facts about the sea-ice cover: Sea ice moves. As with the ice sheets, the ice does not simply form and melt in place; rather, it is in a constant state of motion. Indeed, it moves much faster than do the continental ice sheets. The typical pattern of sea-ice drift in the Arctic Ocean is shown in Figure 6-11. The map reveals two primary features of the ice circulation: a clockwise circulation, or gyre (the Beaufort Sea Gyre), in the west, and the Transpolar Drift Stream, which flows across the pole from the Siberian coast to the East Greenland Sea. This ice flows into the East Greenland Current, where it is transported southward out of the Arctic Ocean. On average it takes about 5 years for ice that forms off the Siberian coast to be transported across the Arctic Basin. Ice that forms in the Beaufort Gyre, however, may last much longer, possibly circling the western Arctic several times before melting in the summer or becoming caught in the Transpolar Drift.

Moved around by winds and ocean currents, the ice is broken into individual pieces (ice floes) that continuously join and break up. Where floes collide, large mounds of ice are thrust up into pressure ridges; where floes move apart, they produce areas of open water called leads or **polynyii** (singular, polynya). Leads are linear open-water features, and polynyii are irregular areas of open water. In winter, these open-water areas freeze very rapidly, but because the icepack is so dynamic, there is always a small amount of open water present, even in midwinter. This is important for two reasons: One, the presence of open water allows for the constant production of new ice, thus releasing salt to the upper ocean and increasing the density of the surface layer. Two, the open water has a considerable impact on the Arctic energy budget: The ocean loses heat to the atmosphere about 100 times faster from the open water than it does through the insulating ice cover. The combined effect of the positive ice–albedo feedback and the heat loss from the

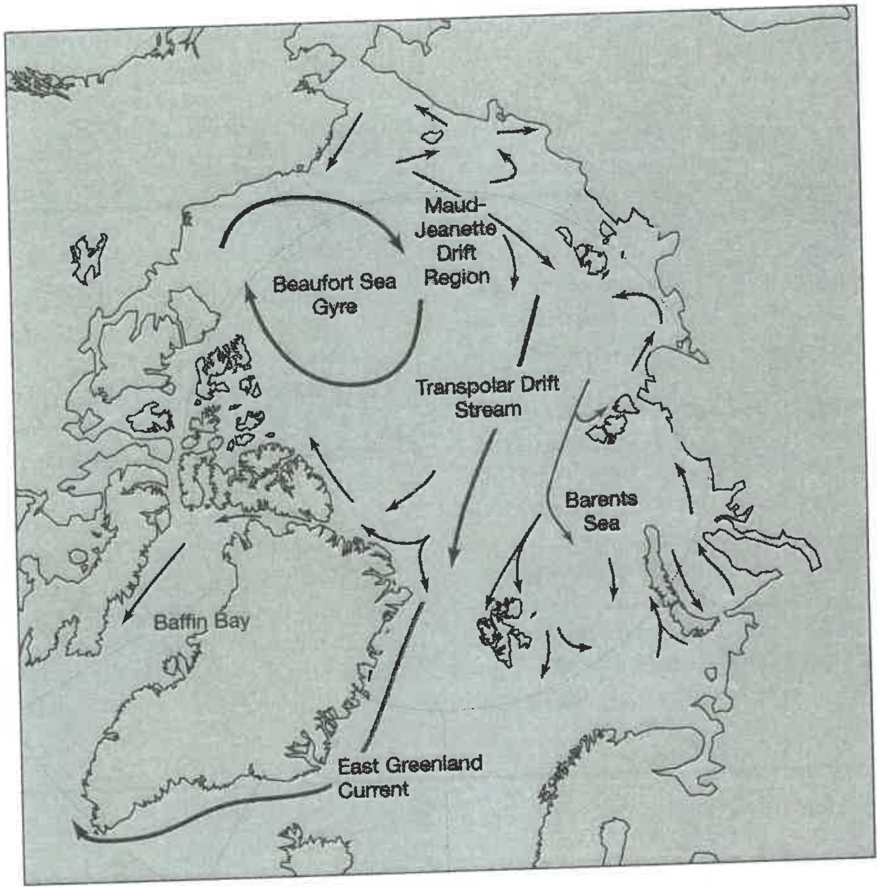


FIGURE 6-11 The typical pattern of sea-ice drift in the Arctic Ocean.

ocean to the atmosphere in areas of low ice concentration both have a significant influence on the high-latitude energy budget. The potential impacts of this become obvious when we look at global model projections of future global warming in Chapters 15 and 16, where these positive feedbacks dominate the high-latitude climate response.

Ice–Atmosphere Interactions

We saw in Chapters 4 and 5 that the sea-ice cover forms as a response to ocean temperatures and that the distribution is then modified by winds and ocean currents. However, the ice cover is not simply a response to the climate: The ice cover modifies the atmospheric and oceanic circulations. We have seen that ice production in the North Atlantic contributes to the formation of North Atlantic Deep Water, which is a major factor in driving the oceanic thermohaline circulation. We have also discussed the ice–albedo feedback, whereby a change in temperature causes a change in the ice cover and the surface albedo, which further modifies the temperature. Recall that this is a positive feedback: Increasing temperature → decreasing ice cover → decreasing albedo → increasing temperature. This feedback can operate year-round near the sunlit ice margins but can play a role only in summer at higher latitudes. (Recall that there is no solar radiation at high latitudes in winter.)

Another sea-ice feedback, however, is important year-round. The ocean surface is a source of thermal energy and latent heat for the polar atmosphere, and the most rapid heat transfer from the ocean to the atmosphere occurs through the areas of thin ice or open water. Thus, we can picture a second positive feedback (Figure 6-12): Increasing temperature leads to decreasing ice cover, which leads to increasing ocean heat flux to the atmosphere, which leads to increasing temperature. Both the ice–albedo feedback and the ocean heat-flux feedback are related to the ice extent and concentration (that is, the area of ice-covered

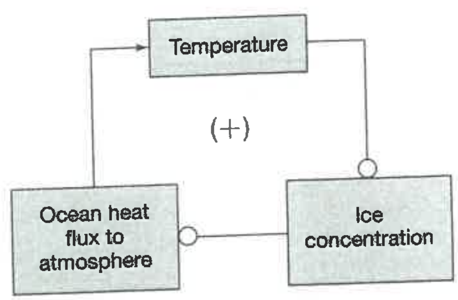


FIGURE 6-12 The sea-ice–ocean heat-flux feedback loop.

ocean). The heat-flux feedback is also controlled to some degree by ice thickness: The heat loss from the ocean to the atmosphere increases as ice thickness decreases. The extent of the ice cover, the ice concentration, and the ice thickness are all determined by the energy budget and the ice dynamics and these, in turn, all act to change the energy balance at the surface and the distribution of energy between the atmosphere and ocean.

Chapter Summary

1. The cryosphere represents the frozen part of the Earth system. It includes glaciers and ice sheets on the land surface, frozen lakes and rivers, sea ice on the oceans, and frozen ground. The continental ice sheets—Greenland and Antarctica—represent the largest sources of freshwater on the planet. If both Antarctica and Greenland were to melt, it would raise global sea level by a little over 60 m.
2. The cryosphere is a dynamic part of the system; both glaciers and sea ice are continuously moving.
   a. Glaciers move under the influence of gravity, with the ice flowing primarily in response to the ice-surface slope, and by sliding over the bedrock.
   b. Sea ice, on the other hand, is carried over the ocean surface through the action of winds and ocean currents, which play a significant role in determining the geographic distributions of sea ice in the Northern Hemisphere.
3. Sea ice, snow cover, and the continental ice sheets all play a significant role in determining the energy balance at high latitudes, primarily through the ice–albedo feedback. This feedback dominates the response of the climate system at high latitudes. For this reason, we would expect some of the most obvious effects of global warming to first appear at high latitudes. During the last decade we have, in fact, seen significant changes in sea-ice extent and melt from the Greenland ice cap.

Key Terms

ablation zone
accumulation zone
active layer
basal shear stress
continental ice sheet
critical point
firn
freezing nuclei
mountain glaciers
normal stress
permafrost
phase diagram
polynyii
sea ice
shear stress
stress
supercooled water
thermal inertia
thermokarst
triple point

Review Questions

1. How does a snow cover affect overlying air temperatures?
2. How does the snow cover affect soil temperatures?
3. How do snow crystals form?
4. What is permafrost?
5. Why might melting permafrost be important for future climate change?
6. Explain how snow is transformed into glacier ice.
7. Explain how glaciers move.
8. Explain the difference between the way glaciers and sea ice accumulate ice.
9. What role do ocean currents play in the distribution of sea ice?
10. What are the two primary ways sea ice affects climate?

Critical-Thinking Problems

1. The first decade of the 21st century has been characterized by significant reductions in the extent of summertime Arctic sea ice. Given what you know of ice–climate interactions and what you learned about the climate system in Chapter 4, explain what effect you think continued decreases in North Atlantic ice cover may have on European climate.
2. Using the density for ice (that you can find in the text) and assuming an ice thickness of 130 m and a surface slope of 5°, calculate the basal shear stress under the glacier. The number you obtain is a typical value for a glacier. If the surface slope changes, the thickness of the ice will change. Assuming that the basal shear stress remains the same, how thick will the ice be if the slope increased to 10°? What does this tell you about the relationship between surface slope and glacier thickness?

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