Chapter 16

The Marine Cryosphere

David M. Holland
Courant Institute of Mathematical Sciences, New York University, New York, USA

Chapter Outline

1. Introduction 413
   1.1. Marine Cryosphere 413
   1.2. Ice Physics 415
   1.3. Ocean Impacts 416
   1.4. Relation to Other Chapters 417
2. Sea Ice 418
   2.1. Observations 419
   2.2. Modeling 420
   2.3. Ocean Mixed-Layer Interaction 421
   2.4. Polynyas 422
   2.5. Impact on Water Masses, and Circulation 422
   2.6. Biogeochemical Ramifications 423
3. Land Ice 423
   3.1. Observations 423
   3.2. Modeling 423
   3.3. Ocean Mixed-Layer Interaction 425
   3.4. Impacts on Water Masses 426
4. Marine Permafrost 428
   5. Emerging Capabilities 429
   5.1. Ice-Capable Observations 429
   5.2. Ocean-Capable Observations 429
   5.3. Ice-Capable Modeling 432
6. Cryospheric Change 432
   6.1. Observed Sea-Ice Change 432
   6.2. Sea-Ice Projections 434
   6.3. Observed Land-Ice Change 435
   6.4. Land-Ice Projections 435
   6.5. Marine Permafrost 436
7. Summary 436

References 437

1. INTRODUCTION

1.1. Marine Cryosphere

The cryosphere is loosely defined as the component of the earth’s climate system consisting of frozen water. In one way or another, all components of the cryosphere interact with and influence the ocean. The cryosphere can be considered to consist of sea ice, land ice, and atmospheric ice. While atmospheric ice, essentially consisting of frozen water clouds, has an indirect impact on the ocean through precipitation, including snowfall onto sea ice and land ice, and atmospheric radiation influences, we discuss it no further here as it is more of an atmospheric science phenomenon. We focus on ice that is more or less in direct contact with the ocean (see Figure 16.1). Land ice, often referred to as a glacier, is in direct contact with the ocean wherever it flows, under the action of gravity, into the ocean to form floating ice shelves along the periphery of ice sheets and ice caps. A particular type of land ice, marine permafrost, is also found beneath the seafloor, where it holds a potential for interaction with the ocean. Sea ice, by its very nature, is formed at the surface of the ocean and is perhaps the most well understood and studied aspect of ice–ocean interaction. Depending upon the time of year, up to 5% of the global ocean surface is covered by ice (NSIDC, 2007). Collectively, we refer to the sea ice, glacier, and permafrost elements as the marine cryosphere.

The interactions of sea ice, land ice, and marine permafrost with the oceans are largely confined to the polar oceans, with the only exception being some marine terminating glaciers found south of the Arctic Circle in, for instance, Alaska (Meier et al., 1980), and north of the Antarctic Circle in Chile (Warren, 1993). The bathymetric layouts of the far northern and southern oceanographic basins are essentially morphological opposites. Looking to the north, the Arctic basin is largely landlocked, the only exchanges with the rest of the global ocean occurring with the North Atlantic Ocean through the Fram Strait and Barents Sea passages, as well as the modest connectivity...
to the Pacific Ocean through Bering Strait, and some exchange through the Canadian Arctic Archipelago. This geographic restriction makes it somewhat easier to observe the water mass exchanges in the north as compared with the south (see Chapter 17). The bathymetry of the Southern Ocean is completely open to the global ocean, with no restriction to exchange with the neighboring South Atlantic, South Pacific, and Indian Oceans (see Chapter 18).

The difference in land configuration in the north and south is a key factor in the distinction between the sea-ice and land-ice cover of the two regions, discussed at length in Sections 2 and 3, respectively. The Arctic Ocean, consisting principally of the area north of 80°N, is surrounded by land that is largely nonglaciated and has river runoff at the surface in the boreal summer that remains largely trapped in the Arctic basin leading to a relatively stratified ocean and thick sea-ice cover. The Southern Ocean, essentially in an unconfined basin and with almost no surface runoff of freshwater, is more weakly stratified and has a much thinner sea-ice cover compared with the Arctic. In other words, there is not really an “Antarctic Ocean” analog to the Arctic Ocean; instead, there exists the Southern Ocean, much of which is subpolar and cannot be compared in any real sense to the truly polar Arctic Ocean.

There are a number of distinctions to be drawn between the coastal and the continental shelf zones of the two polar regions. The Arctic continental shelves cover approximately 10% of the area of the Arctic Ocean and have an average depth of little more than 100 m (Jakobsson, 2002). The situation in the south is fundamentally different with the continental shelf around Antarctica occupying a much smaller percentage area of the Southern Ocean and the average depth of the continental shelf being far greater, on average 600 m (Anderson et al., 1980). The global mean continental shelf depth is estimated to be 150 m (Bott, 1971), and thus, the Antarctic shelf is often referred to as being overdeepened. The deepness is a result of cumulative, erosive activities of a grounded ice sheet (SCAR, 1997). This distinction in average depth is of fundamental importance to the manner in which warm subpolar waters can or cannot go aboard the continental shelves and reach the periphery of the major ice sheets. Another distinct feature of the Antarctic continental shelf is that it is foredeepened, that is, the shelf depth becomes deeper as one travels from the shelf break in toward the margin of the ice sheet (Pope and Anderson, 1992).

The modern marine cryosphere of the north consists of perennial Arctic sea-ice cover, the Greenland Ice Sheet, and marine permafrost; in the south, there is the seasonal sea-ice cover and the more massive Antarctic Ice Sheet. On longer timescales, back through the Pleistocene, the northern cryosphere has undergone larger areal changes as ice sheets occupied the Canadian (Broecker et al., 1989) and Scandinavian (Rinterknecht et al., 2006) land masses more often than not during that epoch.

The character of the seasonal and perennial sea-ice cover of the polar seas has been well noted by voyagers dating back over the past millennium, certainly as early as the time of the Vikings in the north (Ogilvie et al., 2000). In more recent times, over the past century or so, explorers and oceanographers have set sail deep into both the Arctic Ocean (Nansen, 1902) and the Southern Ocean to the coast of the Antarctic continent (Ross, 1847). The mere existence of sea ice, aside from its physical implications for ocean circulation in the high latitudes, has in itself greatly hampered the exploration and understanding of polar ocean dynamics. The difficulty of cutting through a meter or more of frozen ocean surface and the associated risks of being trapped in the “pack” for a year or more have brought many an expedition to a tragic end (e.g., the Franklin expedition; Berton, 1988) or at least a difficult ending (e.g., Amundsen’s travel through the Northwest Passage; Berton, 1988). In the 1890s, Nansen deliberately set his ship into the Arctic ice pack on the Siberian coast,
and riding with the transpolar drift, he crossed close to the North Pole, ultimately to emerge into the Fram Strait area, thereby demonstrating the basic circulation pattern of Arctic sea ice (Nansen, 1902). In a somewhat analogous, albeit not so planned adventure, Ernest Shackleton was set into the pack of the Weddell Sea, and through the trajectory of his ship and travels, the circulation pattern of the ice of the Weddell Sea was better established (Shackleton, 1919).

While sea ice is an evident participant in the marine cryosphere, perhaps not so obvious is the role of land ice. Both the continent of Antarctica (Drewry et al., 1983) and the island of Greenland (Bamber et al., 2001) are almost everywhere covered by an ice sheet several kilometers in thickness. In some locations, these ice sheets actually sit upon bedrock that is below modern-day sea level. Such areas, referred to as marine ice sheets (Mercer, 1978), have direct interaction with the ocean. In some instances, the marine ice sheets terminate at the ocean in vast extensive floating ice shelves, such as in the Ross (Shabtaie and Bentley, 1987) and Weddell seas (Wainwright et al., 1988), while in other cases, the termination can be rather more abrupt with the glacier only presenting a rather modest-sized interface to the ocean, as in the case of many outlet glaciers in Greenland (Cathal et al., 1999). Large portions of the West Antarctic Ice Sheet were discovered to be well below sea level during expeditions undertaken in the International Geophysical Year (Bentley and Oosten, 1961). Such marine-based glaciated areas can be thought of as being an extension of the global ocean. There is some evidence from seafloor sediment records that the ocean has occupied these areas from time to time during past interglacial periods (Scherer et al., 1998). It is the waxing and waning of the marine ice sheets that holds the greatest present interest for potential rapid global sea-level change in the current century and beyond (Alley et al., 2005).

1.2. Ice Physics

To better understand the manner in which ice interacts with the ocean, we first review a few of the fundamental physical properties of ice and freezing ocean water. The water molecule, H$_2$O, has more solid phases than any other known substance. The most common form of solid ice takes on a hexagonal pattern, termed hexagonal ice, or ice Ih, or simply ice. Such ice, created from water at temperatures below 0 °C, settles into a regular crystalline structure with hexagonal symmetry (Hobbs, 1971). At even lower temperatures, ice crystallizes into structures with cubic symmetry, but we are only concerned in an oceanographic context with the hexagonal form. One exception is clathrates, discussed in Section 4 in the context of marine permafrost.

A phase diagram for water, showing the occurrence of the liquid, solid, and gas phases as functions of ambient temperature and pressure, reveals that the melting line has a negative slope (Figure 16.2). This means that as the pressure on ice is increased it tends to move from the solid toward the liquid phase. This is a consequence of ice expanding as it solidifies, giving rise to the remarkable fact that the water solid phase (H) is less dense than its liquid phase, so that ice shelves, icebergs, and sea ice all float upon the ocean, rather than resting on the seafloor, as would be the case if the ocean were composed of almost any other liquid substance.

The mechanical strength of ice varies depending upon the scale under consideration. At the smallest scale, that is of an individual floe of sea ice or a small iceberg, ice is a rigid crystalline solid, gaining its strength from hydrogen bonding, and it fractures in a brittle fashion (Petrovic, 2003). By contrast, on the large scale, sea ice (Mellor, 1983) and ice sheets (Weertman, 1983) follow distinctly different flow regimes than at the small scale. We will further investigate these large-scale dynamical regimes in Sections 4.2 and 4.3 as they are of great consequence to understanding the impact of ice upon the ocean. Deeper insight into the small- versus larger-scale behavior of ice comes from recognizing that, at the level of the crystal structure, individual molecules are linked together in a single crystal, but at the larger scale, ice is a polycrystalline material and its strength is dependent on that structure (Hooke, 1981). Ice has both an elastic behavior, arising from small displacements from their equilibrium position of atoms in a crystal, and a plastic behavior which emerges from the motion of dislocations that are the defects in the otherwise perfect crystal structure. An in-depth treatment of the molecular and aggregate strength properties of ice is found in the textbook by Petrenko and Whitworth (1998).

The thermodynamical properties of ice govern its growth and decay. One has to consider the density, specific heat, latent heat, and related thermodynamical properties of ice to model its evolution in a given thermodynamic regime. For sea ice, which is marine in origin, the thermodynamics can be complicated by the presence of significant amounts of brine locked into the ice (Cox and Weeks, 1982), something not found in glacier ice which is generally meteoric in origin. Whereas freshwater reaches maximum density at a temperature of approximately 4 °C, the impact of salinity in typical seawater is to move the temperature of maximum density to the freezing point (Fujino et al., 1974). A thorough review of ice thermodynamics is provided in the textbook by Lock (1990), which delves into the thermodynamical Gibbs function, which allows one to develop the phase diagram for ice, and thus to understand the manner in which water switches between its phases of liquid, gas, and solid (see also Chapter 6).

The freezing point of water is dependent upon both pressure and salinity (Fujino et al., 1974). The details of these dependencies have a fundamental impact on ice formation and decay in the ocean. The near-ubiquitous...
presence of salt in the ocean lowers the freezing point below the nominal 0 °C mark of pure water, reaching down to −2 °C for typical ocean salinity. The deeper a piece of ice is found in the ocean, the greater the pressure and the lower the freezing point yet again. Combined, the effects of salinity and pressure can reduce the freezing point for some deep glaciers to below −3 °C. The effect of the depth dependence on the freezing temperature can result in more rapid melting for the deeper parts of a glacier than for its shallower parts.

Salinity plays a crucial role in the melting of ice for yet another reason. The molecular diffusivity of salt is approximately one hundred times smaller than that of heat, and as a result, the transfer of heat at the interface where ice and water meet is much more efficient than that of salt. This is because turbulence is suppressed in the upper few millimeters of the boundary layer so that exchanges between ice and ocean occur only by molecular diffusion (Josberger and Martin, 1981). This diffusivity contrast has the effect of making the interface less salty than the ambient fluid during melting, leading to rising of the freezing point, and ultimately slowing of the rate of ice melt compared to the rate that would occur if such an effect were absent (Holland and Jenkins, 1999). In the case of freezing, the situation is further complicated by the presence of convection at the interface due to the unstable fluid stratification that occurs when ocean water freezes and salt is rejected, by and large, from the newly forming ice.

1.3. Ocean Impacts

The presence of an ice cover, whether sea ice or ice shelf, has a profound effect on the underlying upper ocean. Perhaps most notable is the transformation of a low-albedo surface (typical ocean value of 0.10) to a high-albedo surface (typical snow cover surface value of 0.90). The abrupt decrease in transmission and increase in reflection of shortwave radiation as an open-ocean area is converted to a sea-ice-covered area completely alter the surface energy balance (Allison et al., 1993). This gives rise to the so-called ice-albedo effect whereby an increase in sea-ice cover leads to less surface absorption of heat energy, which in turn promotes further growth of sea ice and further decrease in surface energy absorption (Robock, 1980). In a paleoclimate context, such an effect has been widely implicated in the “snowball earth” phenomenon (Hoffman and Schrag, 2002) whereby almost the entire global ocean surface is covered by ice. The white surfaced planet earth absorbs little surface radiation in such a scenario, thereby locking the planet into an icy state. The converse, a planet with no ice cover absorbs an enormous amount of radiation, allowing it to stay in an ice-free state. Likely, a bifurcation
point exists in the earth’s climate system allowing the earth to toggle between ice-covered and near-ice-free states, like the present (North et al., 1981).

Aside from its impact as a radiative barrier, the sea-ice cover is a highly effective barrier to gas exchange between ice and ocean. A key greenhouse gas, carbon dioxide, is continuously exchanged between the ocean and atmosphere, the rate of exchange being controlled by the partial pressure of gas in the atmosphere and the wind speed, and the temperature and salinity of the ocean, which controls the solubility of the gas in seawater (Wanninkhof, 1992) (see also Chapter 30). It is known from observational data that this exchange is highly seasonal dependent, demonstrating the key moderating role played of the ice cover (Golubev et al., 2006). In ice-shelf-covered portions of the ocean, there is no atmospheric-ocean exchange of gas, but there is input of gas into the ocean where the base of an ice shelf melts, as gases dissolved in the ice are released into the ocean, providing a glacial source of dissolved oxygen (Rodehacke et al., 2007).

Throughout the global ocean, the cycle of surface evaporation and precipitation, driven by the atmospheric state and ocean surface conditions, ice-covered or otherwise, helps set the salinity of the ocean surface. Salinity is the determining factor on ocean density in cold waters in part because the thermal expansion coefficient of water decreases with decreasing temperature, hence the importance of surface freshwater fluxes at high latitudes in controlling the downwelling branches of the global thermohaline circulation (THC), also known as the global conveyor belt (Manabe and Stouffer, 1995). The formation of sea ice at high latitudes results in the injection of salt into the surface ocean, often resulting in unstable stratification leading to deep or bottom convection (Aagaard and Carmack, 1989), enhancing the conveyor belt. On the flip side, the melting of sea ice along the sea-ice margin can lead to an input of freshwater that can suppress or even shut down convection, thereby slowing down the conveyor-belt transport of heat from low latitudes to high latitudes.

Particular to the Arctic is the input of substantial river runoff mentioned earlier (~3500 km$^3$/year, Lammers et al., 2001); in fact, some 10% of the global river runoff goes into the Arctic basin (Peterson et al., 2002). The arrangement of basin layout and river input in the Arctic leads to the strongest halocline to be found anywhere in the global ocean (Nguyen et al., 2012).

The strong halocline effectively insulates the Arctic sea-ice cover from the vast, latent heat supply of the Atlantic layer below and thereby allows a substantial thickness, at least historically, of sea-ice cover to persist in the Arctic Ocean (Steele and Boyd, 1998). By contrast, the Southern Ocean has almost no surface freshwater runoff from the Antarctic continent, and the water column is at best marginally stable and thus more easily allows for the upwelling of deep, warm water (Foldvik and Gammelsrød, 1988) that can contribute to a thinner sea-ice cover. The melting at the base of the ice shelves and calving of icebergs cannot really be thought of as a kind of “runoff,” as the melting related to those processes (~2400 km$^3$/year; Rignot et al., 2013) occurs largely at depth within the ocean. The resulting, diluted meltwater contributions are evidently not large enough in quantity to enforce substantive stratification in the surrounding ocean. Massive, persistent convective events, such as the Weddell Polynya event of the mid 1970s, are indicative of this borderline stability in the south (Holland, 2001).

Throughout most of the global ocean, the input of turbulent kinetic energy into the surface ocean leads to a downward mixing of surface properties over a depth scale referred to as the mixed-layer depth (de Boyer Montegut et al., 2004). This depth represents a balance between buoyancy input at the surface ocean, which tends to thin the mixed layer, and mechanical stirring, which tends to deepen the layer. Over a sea-ice-covered ocean, as compared to the open ocean, this balance is altered both in terms of buoyancy and mechanical input details, but the same fundamental principles apply (Lemke et al., 1990).

The mixed layer beneath an ice shelf is another distinct environment for mixed-layer physics (Holland and Jenkins, 2001). Whereas sea ice does move with respect to the ocean surface, an ice-shelf base does not, and consequently, there is much less mechanical input into the mixed layer at the base of an ice shelf. That notwithstanding, the base of an ice shelf tends to be inclined somewhat and thus an effective gravity current can flow upward along the basal slope, leading to a unique source of mechanical mixing, different from that beneath sea ice, which tends to be relatively flat on such scales. A one-dimensional numerical model of a plume flowing up the base of an ice shelf has been provided by Jenkins (1991).

The final impact of ice on the ocean that we mention is that of global sea-level change. Ice floating upon the ocean, such as sea ice or ice shelves, makes only a minor contribution to sea-level change as the ice upon melting freshens the relatively salty ocean, thereby raising the sea level through modest mass contributions (Jenkins and Holland, 2007). More impressive, land ice grounded on bedrock below current-day sea level has a volume above flotation (i.e., the volume of grounded ice above and beyond that which is simply displacing ocean water) that can contribute substantially to global sea-level change. Marine-based West Antarctica has been estimated to hold more than 3 m of potential global sea-level change (Bamber et al., 2009).

1.4. Relation to Other Chapters

This chapter, while dealing primarily with the physics and phenomenology of sea ice and land ice, implicitly draws in
material that is discussed elsewhere in this book, such as mixed-layer dynamics in relation to sea-ice cover and bottom water formation in relation to sea-ice growth and land-ice basal melting. The state of sea ice and land ice in the polar regions is greatly influenced by oceanic heat transport from the subpolar oceans into the polar (see Chapter 11). The freshwater produced by melting sea ice and land ice may exert influence on the overturning circulation in both the Nordic Seas (see Chapter 17) and the southern boundary of the Southern Ocean (see Chapter 18), particularly in the deepwater formation regions of the Ross and Weddell Seas.

The outline of the remainder of this chapter is as follows. Sea ice is discussed in Section 2, followed by land ice in Section 3, and a brief mention of marine permafrost in Section 4. Recent advances in observations and modeling capabilities both for sea ice and land ice, as well as of the waters of the polar oceans, are treated in Section 5, which covers emerging capabilities. Finally, a review of documented, recent changes in the marine cryosphere and an outlook for the twenty-first century are given in Section 6.

2. SEA ICE

The surfaces of the polar oceans have a net energy loss through the course of an annual cycle. An important consequence is that the surface of the ocean freezes over, forming sea ice, and as solid ice is less dense than liquid water, the ice floats upon the ocean surface, turning it from dark blue to bright white. With the swing of the seasons alternating between the northern and the southern hemisphere, the sea-ice cover of the Arctic and Antarctic grow and retreat out of phase with one another, producing one of the largest seasonal phenomena on the earth’s surface (see Figure 16.3). The Arctic sea-ice cover varies from a wintertime maximum of $15 \times 10^6 \text{ km}^2$ to a summertime minimum of 7, whereas the Antarctic swings between 20 and 4 (Gloersen et al., 1992).

The detailed process by which sea ice forms depends on the state of the ocean’s surface, as a more turbulent ocean will lead to a somewhat different ice cover than a calm ocean. In either case, atmospheric cooling of the ocean results in its surface being slightly supercooled in the first instance, which in turn initiates nucleation of ice crystals and frazil ice. The frazil ice platelets coalesce, leading to a type of ice known as nilas in relatively calm seas and to pancake ice in more turbulent seas, where distinct floes are discernible (ASPECT, 2012). Floes can be pushed to ride up upon one another, thus thickening the cover. Once a sufficient cover exists, additional growth can occur at the underside of the ice, whereby loss of heat by conduction through the cover leads to formation of congelation ice directly to the underside.

Growth of ice can also occur at the ice surface. This happens when snow cover accumulates on the ice surface to the extent that the weight of the snow is sufficient to depress the freeboard of the ice below sea level, causing flooding of the ice surface and conversion of wet snow to ice, the resulting ice being referred to as snow ice (Jeffries et al., 1997).

Ice that forms from the sea, that is, ice of marine origin, has a significant salt content, at least initially. The salinity of the ocean is approximately 35 psu, and when frazil ice forms, this salt is rejected. However, the rapid formation of sea ice leads to some salt being trapped into pockets within the larger sea-ice conglomerate, giving the ice a bulk salinity of approximately 3–4 psu (Niedrauer and Martin, 1979). These distributed pockets become very concentrated in salt and do not freeze. Over time, however, they tend to drain down through the ice under the action of gravity. Melt water washing through the ice during summertime as the result of surface melt pond formation is another factor that gradually reduces the salt content of sea ice (Fetterer and Untersteiner, 1998). Glacier ice, by contrast, is generally of meteoric origin and has very limited salt content, which is largely aeolian in origin.

A fully developed ice cover is usually relatively mobile, blown around by the winds and dragged by the ocean currents. Most of the Arctic and Antarctic sea-ice cover is mobile, but there are areas near the coast where the ice is stagnant and referred to as landfast ice (Mahoney et al., 2007). The main characteristics of an ice cover are its thickness and concentration. The thickness of Arctic sea ice ranges from thin, that is, 0.1 m, to thick, that is, up to 6 m, whereas Antarctic sea ice tends to be relatively thin overall, generally spanning the range 0.1–1 m. In some locations, the sea ice can pile up substantially upon itself, forming ridges that can reach up to 20 m in thickness. The term “concentration” is used to describe the fractional areal coverage of the ocean surface by sea ice and ranges between 0% and 100%. While in the case of landfast ice the coverage is full, 100%, the majority of the ice cover has leads, or fracture zones, in which the ocean surface is directly exposed to the polar atmosphere. Even though the leads occupy only a few percent of the sea-ice region, it is because of the great efficiency with which they can exchange heat and moisture with the atmosphere that they are an important aspect of the ice cover.

A major distinction between sea ice of the north and that of the south is the age of the ice. Arctic ice can have a significant multiyear fraction, meaning that the ice has survived several seasons—largely the result of the enclosed nature of the Arctic basin. By contrast, Antarctic ice is almost exclusively first-year ice as each summer season the ice cover is largely melted away. A similarity of Arctic and Antarctic ice is that they both have marginal ice zones, that is, transition zones between the interior, densely concentrated pack ice and the ice-free open ocean. The marginal ice zone migrates outward during winter and
retreats inward during summer, giving rise to an area of coverage known as the seasonal ice zone.

2.1. Observations

The traditional observational strategy for sea ice has been to make observations from ships traveling near or through the pack. Much of this data is dependent upon the subjective evaluations of an onboard ice observer, leading to some concerns over data quality. An effort has been made to standardize the observations, leading to a more quantitative global data set (ASPECT, 2012). From a ship, estimates of sea-ice concentration, often quoted in “egg-chart” fashion as fractions of eight, and sea-ice thickness are feasible.

While a single ship-based cruise provides a relatively small areal data sample, taken cumulatively over many cruises, the global data set is of value to sea-ice researchers.

Complementing the ship-based data set is a submarine observational data set, based primarily on upward looking sonar that can give an estimate of ice areal coverage and draft (Moritz and Wensnahan, 2011). The draft can be converted to ice thickness, subject to assumptions about the ice density and snow cover. The data can be used to evaluate decadal and multidecadal trends in sea-ice thickness, spanning back to 1958 (Kwok and Rothrock, 2009). Presently, this type of data only exists for the Arctic region.

Another approach used to collect sea-ice data along a track has been the development and deployment of buoys.

---

**FIGURE 16.3** Seasonal cycle of sea-ice cover in both hemispheres. Top panels: Arctic winter maximum in March and summer minimum in September. Bottom panels: Antarctic summer minimum in March and winter maximum in September. Blue areas are open ocean, gray areas are continental land mass, and white is sea ice. [http://nsidc.org/cryosphere/sotc/sea_ice.html](http://nsidc.org/cryosphere/sotc/sea_ice.html).
placed on the surface of the ice, both in the Arctic (IABP, 2012) and the Antarctic (IPAB, 2012). The original version of these buoys collected surface air temperature and sea-level pressure data, and reported their position, all highly valuable as input to weather forecasting models and subsequent atmospheric reanalysis products. The buoys provided a very clear view of the circulation pattern of Arctic and Antarctic sea ice and the relation of that circulation to the sea-level pressure fields, and hence wind patterns. More recent development in sea-ice buoys now has them placed on the ice surface but with instrumentation extending all the way through the ice into the top few meters of the ocean. The buoys now have the ability to measure the temperature through the sea ice as well as the ice thickness and the entire mass balance of an ice floe. Such buoys have been developed that can not only be deployed in ice but also float on the ocean in the instance that the ice floe melts entirely (SIMB, 2012).

The largest leap forward in sea-ice observations occurred in the 1970s with the deployment of microwave satellite remote sensing instrumentation (Gloersen et al., 1992). The data provided the first global view of the entire Arctic and Antarctic sea-ice fields, their seasonal variability, and their interannual trends. Data sets were limited initially to primarily sea-ice concentration and used a variety of techniques to convert brightness radiance to concentration, producing broadly similar data products. More recently, developments using particle tracking, or image correlation techniques, with data sets from visible or active microwave sensors have allowed sea-ice velocity fields to be assessed, giving an impressive view of the large gyres that exist with the ice packs (Vesecky et al., 1988). Most recently, laser altimetry has provided the first glimpses of sea-ice thickness, based on measurements of the freeboard of ice (Hvidegaard and Forsberg, 2002).

2.2. Modeling

To understand the physics of large-scale sea-ice behavior, a model is needed, theoretical in nature and numerical in implementation. From a pragmatic view point, a numerical model once developed can be used to make projections of sea-ice behavior. Sea-ice physics, on the large scale, is complex, even more so than the ocean in at least one respect—while the continuum hypothesis upon which ocean modeling is based has some reasonable foundation, it is not the case for sea ice. Sea ice is composed of individual floes at scales ranging from 1 km down to 1 m, making the description of the sea-ice fluid exceptionally challenging. That notwithstanding, parameterizations of sea ice allow one to describe in a model the aggregate behavior of the underresolved smaller-scale features. Based on comparisons of model simulations with observational data, it can be claimed that there is indeed some accuracy and skill in the current generation of large-scale sea-ice models (Fichefet et al., 2003). The early-generation sea-ice models were “stand-alone” in the sense that the atmospheric and oceanic boundary conditions were fed to the sea-ice model equations as fixed boundary conditions (Flato and Hibler, 1992) as opposed to the present generation where the boundary conditions are dynamical, coming from atmosphere and ocean models in which the sea-ice model is coupled. Subsequent development of coupling strategies has now led to the point where sea-ice models are fully coupled with ocean and atmospheric general circulation models used in climate-change simulation scenarios (IPCC, 2007). We next discuss the major conceptual components of a sea-ice model: ice thermodynamics (growth/decay) and dynamics (motion).

Sea-ice thermodynamical equations describe the flow of heat through the sea-ice medium (Maykut and Untersteiner, 1971). As the aspect ratio of ice is such that ice is thin, the flow direction of heat is largely vertical. Brine pockets aside, the flow of heat is governed by a one-dimensional (vertical) equation describing diffusion of heat by molecular processes, aided by the fact that the thermal conductivity of sea ice is a well-established quantity (Ebert and Curry, 1993). The addition of a snow cover requires only a step change in the thermal conductivity to a smaller value (as snow is more insulating than ice) (Grenfell, 1991). The main boundary conditions are that of the flux of heat applied at the base of the sea ice and at its upper surface. Here we describe the surface fluxes and defer the basal fluxes to the mixed-layer discussion below.

At the surface of the ice (or snow), there are two fluxes to consider, the turbulent and the radiative. The turbulent fluxes in turn divide into two categories: the latent heat flux associated with evaporation from leads or sublimation from the ice (snow) surface and the sensible heat flux due to convective heat transfer over the leads or ice (snow) surface. The turbulent fluxes are calculated using bulk parameterizations that seek to capture the unresolved role of turbulence in moisture and heat transfer though empirical relations based on the time-averaged (i.e., bulk) air flow properties. The radiative fluxes also split into two subcategories: the longwave and the shortwave radiation. The simulated upwelling and downwelling components of the shortwave radiation are critically dependent on the description of the surface albedo. The sum of these four fluxes leads to a surface heat source or sink for the sea ice, helping to drive either ice melt at the surface or ice growth at the base (Oberhuber et al., 1993).

A unique aspect of sea-ice modeling, as compared with land-ice or even ocean modeling, is the treatment of the mass-conservation equation. While ice is considered incompressible from a density point of view, similar to the approximation made in ocean modeling, the three-dimensional volume of ice is actually split into two separate
equations: an equation for ice thickness (i.e., the vertical average over an area) and an equation for the areal fraction, the area actually covered by sea ice (with the other fraction denoted as lead area). Taking the combined product of the ice thickness and areal concentration equation allows one to reconstitute the original three-dimensional mass (volume) conservation equation. This mathematical splitting of the volume (mass) conservation equation into separate equations for thickness and concentration is done as it is natural to model sea-ice thickness and concentration as separate variables and also to ease comparisons with observations. While this is a rather clever split, one complication that arises is how to apply heat sources and sinks to ice growth and decay. That is, just as the volume equation has been split into two equations, the heat source/sink terms need also to be split, but it is not at all obvious how to partition the heat source/sink. Specifically, given a unit of heat loss from the ocean, should that heat be used to grow the ice vertically and extend the thickness or horizontally and extend the areal coverage? The inverse question can be asked in the case of a unit of heat gain. Current-generation models have a relatively ad hoc formulation of how to split this heat flux, albeit one that generates reasonable model simulations. Further research is required to better understand and parameterize this heat flux split.

The sea-ice dynamics equations describe how the pack ice moves under the influence of various forces: ocean surface tilt, Coriolis effect, wind, currents, and most interestingly, internal ice dynamics, often referred to as ice rheology. The main discussion in theoretical developments of sea ice centers on an appropriate rheology—or flow law—that accurately describes how ice floes interact with one another and how they exchange momentum. Ocean waters are treated as a Newtonian fluid whereby stress is linearly related to strain through a fixed viscous coefficient. By contrast, sea ice is thought of as an elastic–plastic fluid (Rothrock, 1975). At small strain, the stress is described as elastic and is reversible. At larger strain, the stress is finite and plastic, meaning the ice fails under tension, compression, or shear stress. An appropriate description, one that is both physically accurate and numerically efficient, has been a target of sea-ice dynamicists for decades. Progress has been made, with the current generation of models simulating ice quite reasonably, starting notably from the early work of Hibler (1979) in which the observed sea-ice thickness pattern of the Arctic Ocean, with thick 6-m ice on the Canadian side and thin 1-m ice on the Russian side, was reasonably simulated for the first time.

2.3. Ocean Mixed-Layer Interaction

The field of oceanography took a major theoretical step forward with the discovery of the phenomenon of the Ekman layer (Ekman, 1905). During his trans-Arctic cruise, Nansen noticed that sea ice did not flow in the direction of the prevailing wind but drifted somewhat to the right. Nansen had this observation passed to Ekman who subsequently worked on the mathematical details of how sea ice and the upper ocean interact. Ekman’s mathematics described a layer beneath the sea ice that progressively veered the water current to the right in a spiral fashion, with deeper water moving slower due to viscous forces. The underlying dynamics is a simple balance between Coriolis force and vertical stress. The resulting spiral structure has been directly observed beneath sea ice (Hunkins, 1966), as well as at the open ocean surface (England et al., 1993) and at the seafloor (Kundu, 1976). The Ekman layer is of significant dynamical consequence as it drives a net transport of water to the right (left) of the surface ocean stress in the northern (southern) hemisphere. In the context of the marine cryosphere, this has an important role to play in the upwelling or downwelling of waters along the margins of the Arctic and Antarctic (England et al., 1993).

Away from the sea-ice-covered portions of the global ocean, the surface water layer is mixed down to some depth, generally in the range 10–100 m. In the resulting surface, mixed-layer water properties are almost homogeneous in the vertical (de Boyer Montegut et al., 2004). The key principle underpinning this mixed-layer depth is the balance between turbulent kinetic energy input and its ability to overcome the potential energy of the ambient stratification. The turbulent input comes via surface wind stress, ocean current shear, and wave breaking, and the opposing potential energy barrier is simply due to the ambient stratification (Large et al., 1994). In the Arctic Ocean, one of the most stratified in the world, the top 100–200 m is well stratified with respect to the deeper layers of water, making exchange between the upper, cold halocline and the deeper, warmer Atlantic layer quite difficult from an energy point of view (Aagaard and Carmack, 1989). By contrast, the Southern Ocean is only weakly stratified and a modest input of turbulent energy can overcome the weak background stratification (Martinson, 1990).

Vigorous exchange in the upper ocean, driven by turbulence, rapidly homogenizes scalar properties. In the case of a sea-ice cover, the exchange between the liquid ocean and solid ice surface, specifically at the liquid–solid interface, occurs via molecular transfer processes. The underlying principle is that turbulent fluid exchange is dampened at a solid interface as the scales of turbulence decay as one approaches the solid interface (Kader and Yaglom, 1972). Extremely close to the interface, the final exchange of scale properties occurs via molecular (diffusive) processes. At the ice–ocean interface, both salinity and temperature play a central role in the melt (growth) rates at the interface. Furthermore, as salt diffusion is approximately 100 times slower than that of temperature, there is an interesting...
competitive dynamic at the interface. The end result is that the ice–ocean interface melt (growth) rate is controlled by the relative rates at which heat and salt can diffuse into the boundary layer. At the interface, there are three physical relations to be satisfied simultaneously: the empirical pressure and salinity-dependent freezing point of sea water, the balance of heat flux into and out of the interface, and the balance between the salt flux into the interface and the freshening effect of melt water from the interface. An illustration of the various scalars and fluxes is presented in Figure 16.4. This set of constraints gives rise to the “three-equation” model of ice–ocean interaction (McPhee, 2008).

2.4. Polynyas

Derived from the Russian word meaning “a large opening in an otherwise sea-ice covered area,” polynyas occur at various locations throughout the wintertime sea-ice cover. Their existence is at first glance surprising, as the surface energy balance in polar winter is strongly negative such that the ocean is losing a vast amount of heat, and one would expect it to be ice covered. Two physical processes that allow for the occurrence of polynyas are, first, the existence of solid boundaries such as coastline which allow ice to be blown away from the coast and slowly replenished by new ice growth (Pease, 1987), and second, vertical convective processes that can raise warm water from depth (Holland, 2001). In either case, adequate heat is brought into the mixed layer or ice transported away from the mixed layer, such that sustained open water areas are possible, even in wintertime conditions.

From the perspective of the climate modeling and physical oceanographic communities, recurrent polynyas are worthy of study (Willmott et al., 2007) because they are sites where (1) water mass transformation takes place through the combined effects of cooling and frazil ice formation, (2) large ocean-to-atmosphere heat (several hundred Watts per square meter) and moisture fluxes occur, and (3) atmospheric CO2 is sequestered into the ocean by physical–chemical processes and biological activity.

In coastal polynyas, where the water is shallow, the ocean quickly cools down to the freezing point at all depths (Lemke, 2001). The heat supplied to the atmosphere then originates only from the latent heat of fusion due to the continuous production of sea ice. Where offshore winds occur, sea ice is constantly formed and blown away from the coast, allowing yet more sea ice to be formed, in effect creating a type of sea-ice factory. During sea-ice formation, oceanic salt is released into the water because it is not incorporated into the ice crystals. This salt increases the local density of the ocean water, in some cases to the point where it can sink to the deep ocean. Coastal polynyas therefore represent important sources of dense deep and bottom water, which ventilate the abyss and drive the oceanic global conveyor belt.

In open ocean polynyas, the supply of oceanic heat is less restricted (Lemke, 2001). Therefore, large amounts of sensible heat from deeper ocean layers are lost to the atmosphere and little ice is created. The substantial cooling of the ocean water can lead to deep convection, which homogenizes the ocean waters to great depths and produces deep water in the open ocean. Consequently, both types of polynyas have a large influence on global ocean circulation and on the earth’s climate.

2.5. Impact on Water Masses, and Circulation

The mere existence of a sea-ice cover provides a distinctive surface boundary forcing compared to other regions of the global ocean. From an oceanographic point of view, the most important implication of the sea-ice cover is the
large input of salt that sea-ice growth inputs to the ocean surface during winter formation. On the flip side, the summer melt of the sea ice provides an enormous freshwater input at the surface. Taken together, these two processes result in an effective salt pump that moves salt from the near surface by convectively driving it down into the deeper ocean (Broecker and Peng, 1987).

A second impact of the ice cover is a modification to the input of wind stress (Joffre, 1982). Well known is that the curl of wind stress is the major input to the Sverdrup balance governing large-scale ocean circulation. Where there is extensive sea-ice cover, particularly near coastal regions, the exchange of momentum between atmosphere and ocean is fundamentally altered by the existence of the sea-ice cover. In the case of a lengthy sea-ice margin, the curl of the wind stress on the ocean can be radically different from anywhere else in the ocean as there is a sharp line where the curl abruptly changes (Quadfasel et al., 1987).

2.6. Biogeochemical Ramifications

Despite the low temperatures of the upper polar oceans, the presence of sea ice provides a platform that allows one of the largest and most important components of the earth’s ecosystem to exist. Despite the low light levels, high salinities, and poor surface gas exchange, microorganisms thrive in this environment, particularly algae (Lizotte, 2001). As the regulation of carbon dioxide between the ocean and the atmosphere is particularly important to the marine biosphere, and sea ice plays a key role in modulating gas exchange and allowing organisms grow, once again the existence of a sea-ice cover is seen to be important in how the global ocean interacts with the atmosphere and ultimately the amount of carbon dioxide stored in the atmosphere, and hence global air temperatures (Rysgaard et al., 2011).

3. LAND ICE

Land ice, built up through the process of evaporation over the global ocean and subsequent precipitation over the great ice sheets of Greenland and Antarctica, has traditionally been thought to be one of the most passive components of the climate system and thus of not much consequence to ocean circulation, particularly on timescales of centuries or less (Figure 16.5). The pace of interaction was considered to be “glacial.” Since the early 1990s, however, an unprecedented rapid change in the margins of the ice sheets has been observed through the eyes of satellites (Nick et al., 2009), and the cause for change has been largely deduced to be the global ocean, via the import of warm waters to the periphery of the ice sheets (Pritchard et al., 2012). Suddenly, land ice has become a major player in rapid global sea-level change and water mass modification processes in the polar oceans (Alley et al., 2005).

3.1. Observations

The global ocean only directly interacts with the ice sheets where the periphery of the ice terminates in the ocean, either as an ice shelf, which extends out onto the ocean surface as a floating tongue, or a tidewater glacier which has an abrupt transition from land ice to ocean with no floating tongue (Meier and Post, 1987). It was once thought that the disintegration of an ice shelf, which by definition floats on the ocean, would have no impact on the flow into the ocean of the inland ice that fed it (Van der Veen, 1985). That viewpoint has been disproved during the past decade as ice shelves, such as the Larsen B Ice Shelf, rapidly disintegrated and the flow toward the ocean of the inland ice behind accelerated manifold (Scambos et al., 2004).

This knowledge comes from an unprecedented suite of observational equipment placed in situ on the periphery of the ice sheets, and airborne and remote sensing instrumentation (Mohr et al., 1998). Remote sensing techniques have shown a doubling in speed of the Jakobshavn glacier in west Greenland, following the breakup of the ice tongue in front of that glacier (Alley et al., 2008). That particular breakup is believed to have been caused by the sudden arrival of warm waters into the fjord connecting to the glacier (Holland et al., 2008). Laser altimetry has captured a drop in surface elevation over the Greenland outlet glaciers, and their analogs in the Antarctic, most notably in the Pine Island Glacier region, located in the Amundsen Sea (Pritchard et al., 2009). At the same time, correlation techniques applied to imagery have demonstrated that the velocity of these outlet glaciers has increased, coincident with the drop in surface elevations (Joughin et al., 2004). Furthermore, the occurrence of simultaneous change between neighboring glaciers suggests the ocean to be the main driver of this change (Shepherd et al., 2004).

3.2. Modeling

The physical basis for modeling land ice holds some similarity with sea ice, described in the previous section, but also has significant differences. A main difference is that glacial ice is treated as a three-dimensional continuum (Blatter et al., 2011), with no analog for the description of concentration (or open space) as is the case for sea ice. Glacial ice does indeed fracture, but such fractures occupy a small portion of the total areal coverage and are not yet widely considered in the overall integrity or strength of an ice sheet or shelf. There are some efforts to correct this modeling deficiency by incorporating the concept of “damage mechanics” into the ice equations (Pralong et al., 2003).

The thermodynamics of glacial ice is analogous to sea ice in almost every respect; that is, boundary conditions apply to the ice surface and base, and molecular diffusion
of heat is the dominant process in the ice interior (Hutter, 1982). Given the thickness of glacial ice, tending to be greater than 1 km, it responds to thermal anomalies on the timescale of hundreds or thousands of years. Sea ice, on the other hand, is typically of order 1 m thick and responds to anomalies at the ice–ocean or atmosphere–ocean boundary rapidly, certainly well less than 1 year.

The dynamics of glacial ice bear little resemblance to those of sea ice, not only because of the absence of fracture or void space in the treatment of glacial ice but also because, unlike sea ice, glacial ice is in general not currently modeled to fracture. This is quite surprising, given the fact that one of the most dramatic natural events to occur is the calving, or fracturing, of large pieces of glacial ice into the coastal ocean at the margins of the great ice sheets. From the coast, the icebergs are transported by the mean ocean currents and winds and subsequently melt into the ocean, providing a distributed source of freshwater into the ocean. In order for this process to be adequately captured in global climate models, significant progress is going to be required in the physical description of ice calving, and hence of ice rheology. This advance is in turn dependent upon the acquisition of adequate observational data to better understand the mechanism underlying ice calving.

FIGURE 16.5 High-altitude views of the two major ice sheets (a) Greenland and (b) Antarctica. The periphery of the Greenland Ice Sheet interacts with the surrounding ocean through relatively narrow, deep fjords, some of which have floating tongues and others which terminate abruptly as a vertical face at the ice–ocean interface. By contrast, the interaction of the Antarctic Ice Sheet with the surrounding ocean is through ice shelves ranging in size from modest to massive, with the three largest being the Ross, Weddell, and Amery. http://www.climatepedia.org/about-ice-sheets; http://news.discovery.com/earth/zooms/ocean-warming-melting-antarctica-ice-sheets-120426.html.
3.3. Ocean Mixed-Layer Interaction

From an oceanographic perspective, the thermodynamic interaction of the base of an ice shelf with the ocean is crucial for two reasons: firstly, the melting process modifies water masses such as Antarctic Bottom Water (AABW) (Foster and Carmack, 1976), and second, the disintegration of an ice shelf could eventually contribute to a change in sea level (Alley et al., 2005). Here we discuss the current treatment of thermodynamic interaction. The physical description is almost identical to that of the analogous phenomena at the base of sea ice, as discussed in Section 2. This is reasonable since the two environments have many features in common.

One important distinction is that the base of an ice shelf can be located at considerably higher pressure than that of sea ice, which is nominally at atmospheric pressure. At depths of 2 km, such as at the grounding zone of the Filchner-Ronne Ice Shelf in West Antarctica, the pressure increase causes a lowering of the freezing point to approximately 1.5 °C below the value at the ocean surface (Fujino et al., 1974).

This has the interesting and important consequence that if water at the surface freezing point (−1.9 °C), the coldest water that can be formed by direct interaction with the atmosphere, reaches ice at the deep grounding zones, it finds itself at least 1.5 °C warmer than the local freezing point and can therefore cause high rates of basal melting.

This thermodynamical effect leads to an ocean dynamical circulation creating the ice pump mechanism (Figure 16.6). Effectively deep ice is melted because of the pressure dependence of freezing temperature. The resulting melt water rises along the base of the ice shelf under the action of gravity because the meltwater is always more buoyant than the ambient—a consequence of the equation of state for seawater at low temperatures. Depending upon the ambient conditions, the rising water may become supercooled as the pressure lessens and the freezing point rises, leading to the possibility that frazil ice may form and be deposited at the base of the ice shelf (Lewis and Perkins, 1986). This entire cycle in which ice is melted at depth and then refrozen at shallower ice-shelf drafts is thought of as an ice pump circulation—moving floating ice from deep to shallow elevations at the ice-shelf base.

In the discussion of sea-ice basal melting in Section 2.3, the concept of the viscous-sublayer model has been introduced (see again Figure 16.4). This model has also been adopted for use at the base of an ice shelf (Holland and Jenkins, 1999), despite not yet being thoroughly validated for use in such a regime. Other than the greater depth, another feature distinguishing a sub-ice-shelf base from a sea-ice base is that an ice-shelf base tends to be inclined, typically by 0.1% slope, but steeper in some locations, particularly near the grounding zone. Field campaigns are only now being carried out to make direct measurements of the turbulent boundary layer at the base of a few Antarctic ice shelves, with validation and improvement of parameterizations of transport of fluxes through the ice shelf–ocean boundary layer likely to occur within the next few years. The first field results suggest that, at least in the instance of the Pine Island Ice Shelf in West Antarctica, the slope of the interface and the stratification that results from the melting along the sloping base form an intense feedback strengthening the melt rates. This feedback likely contributes to the large-scale channelization of the ice-shelf base (Stanton et al., 2013).

We have thus far mentioned facets of the ocean mixed layers beneath a sea-ice cover and also beneath an ice shelf. Along the front of an actively calving glacier, there often exists a sea-ice cover with icebergs embedded, sprinkled through the sea ice. The resulting mélangé of sea ice and iceberg makes for a truly complex ocean mixed-layer environment, one in which the usual turbulent kinetic energy input is also influenced by the stirring due to deep iceberg keels as well as buoyancy input due to freshwater release as the bergs melt at depth (Burton et al., 2012). This is an area of ocean mixed-layer modeling that has not been much addressed to date and one that is in need of in situ observations upon which a solid theoretical foundation can be laid.

While sea ice is an obvious form of ice of marine origin, there is also a form that develops beneath an ice shelf and can reach impressive thickness. For instance, in the middle of the Filchner-Ronne Ice Shelf, the thickness of the bottom-accumulated marine ice layer is estimated to be almost half the total ice-shelf thickness (Oerter et al.,

![FIGURE 16.6 Schematic representation of the two-dimensional circulation under an ice shelf. Salt rejected by winter sea-ice growth forms dense, high-salinity water, which sinks and flows beneath the ice shelf. This causes melt when it comes into contact with deep ice at the grounding zone. The freshened plume rises under the base of the shelf and can either refreeze as marine ice or mix with warm, salty Circumpolar Deep Water (CDW) to form Antarctic Bottom Water (AABW). The circulation is often referred to as an “ice pump.” From AMISR (2013).](image-url)
This marine ice is formed in the water column as part of the ice pump mechanism, mentioned above. The presence of significant thickness of marine ice at the base of an ice shelf could have significant ramifications for the strength of the ice shelf as a whole (Holland et al., 2009) and, if the marine ice survives to the ice front without being remelted, could play an important role in the manner in which an ice-shelf front calves.

### 3.4. Impacts on Water Masses

The continental shelves of Antarctica, particularly the western edges of the Ross and Weddell Seas, are well-known locations for formation of AABW (Foster and Carmack, 1976). This formation is the result of wintertime production of sea ice which causes the surface salinities to reach the highest values found anywhere in the Southern Ocean. As salinity controls density for waters near the freezing point, these cold salty surface waters sink under vertical convection and flow northward down over the continental shelf break and fill the abyssal global ocean. Approximately half of the global ocean volume is constituted by this water mass. But as the continental shelves of Antarctica are foredeepened, some fraction of the dense water formed over the continental shelves first goes southward and into the sub-ice-shelf cavity (see again Figure 16.6). The water that goes beneath the ice shelf generally travels inland all the way back to the grounding zone where, because of the dependence of the melting point of ice on pressure, the newly arrived water causes ice-shelf basal melting. The resulting melt water, known as Ice Shelf Water (ISW), is buoyant and rises along the ice-shelf base as mentioned earlier. In some areas, this ISW plays an important role in AABW production. The production of much of the AABW over the south western Weddell Sea continental shelf is regarded as being a result of ISW mixing with ambient warm, deep water on its way down the continental slope (Seabrooke et al., 1971). In the Ross Sea, some significant proportion of AABW is created by a similar mechanism (Jacobs et al., 1970). In the eastern Weddell Sea, and in many other parts of Antarctica, the melt from ice shelves freshens the continental shelf enough to inhibit the production of high-salinity shelf waters, which in turn inhibits the production of AABW. Turning to sea-ice cover impacts of ISW, it has been shown in numerical experiments that the mere existence of ISW in the Weddell Sea results in a fresher ocean surface over the Weddell Sea and consequently there is an impact on increased sea-ice cover development (Hellmer et al., 2006).

While dense water formation does occur in the Greenland Sea, this is truly open-ocean convection and is not forming along the coast of Greenland where glacier-ocean interaction might have an influence. Instead, it is the melting of the Greenland outlet glaciers by warm waters, from sources distinct from those feeding the convective deep-water formation zones that hold the potential to impact the convective zones (Aagaard and Carmack, 1989). The formation of deep water in the Greenland Sea is a major component of the global conveyor belt, and such formation can in principle be inhibited by the melt water coming from the Greenland ice sheet (Rahmstorf, 1999). In the transition between glacial and interglacial periods, it is thought that the massive freshwater pulse forming from either the Laurentide or the Greenland Ice Sheets feeds freshwater that caps the convective zones (Broecker and Denton, 1990).

### 3.5. Geochemical Tracers

Understanding the basal melting and water mass modifications beneath an ice shelf has been hampered by the difficulty of accessing the cavity. Setting up sustained monitoring of the ice-shelf cavity openings has also been problematic. Aside from traditional CTD (conductivity, temperature, and depth) observations that can be made at such locations either through the ice shelf (Nicholls et al., 1991) or along the ice front (Jacobs et al., 2002), one can also measure the concentration of various geochemical tracers and, to some degree, deduce the amount of glacial input to the ocean arising from basal melt (Smethie and Jacobs, 2005). Additionally, by using measurements of geochemical tracer concentrations to supplement the more common observations of temperature and salinity, one can better constrain the calculation of the amount of mixing between various water masses in the cavity (Schlosser et al., 1990). Another rationale for using geochemical tracers is that temperature and salinity are “nonconservative” when involved in the melting of ice in the sense that latent heat and freshwater exchange occur, transforming the two properties. The use of additional, independent geochemical tracers can remove ambiguity in water mass mixing calculations in such environments.

From an oceanographic point of view, geochemical tracers can be either naturally occurring isotopes of, for instance, oxygen, helium, or neon, or those of anthropogenic origin such as CFCs (chlorofluorocarbons). Over the past half century, CFCs have been entering the atmosphere and therefore the interior of the ocean through gas exchange at its surface (Beining and Roether, 1996). For the purpose of water mass analysis and mixing ratios, if one knows the amount of CFC in parent water masses, then one can better estimate the amount of relative mixing...
between two or more such water masses based on CFC measurements in the mixed-water mass. In the instance of an ice-shelf base, knowing CFC concentration in the high salinity shelf water (HSSW) that forms in the open ocean surface adjacent to the ice front and knowing the CFC concentration in ISW that subsequently forms when HSSW interacts with the base of an ice shelf, one is able to deduce the amount of input meltwater. One complicating factor in the case of the ice-shelf regime is that the presence of a variable sea-ice cover in the formation region means that equilibration between atmosphere and ocean is not always established, making the overall calculations subject to greater uncertainty (Smethie and Jacobs, 2005).

Other than CFCs, oxygen, helium, and neon have also been used to study glacial melt into the ocean. Oxygen, helium, and neon gas are present in glacial ice trapped in air bubbles or as clathrates (see Section 4). These gases are released from the ice and dissolve in seawater as a result of ice-shelf basal melting. The low solubility of helium and neon in seawater results in concentrations well above solubility equilibrium with the atmosphere, producing a glacial melt signal that can be readily traced from an ice front, across the continental shelf and into the abyssal ocean (Schlosser et al., 1990).

Geochemical tracers aside, another powerful constraint on calculating water mass mixing and formation in the polar regions is that relating to the constrained ratio which temperature and salinity change must follow as a water mass gives up heat to melt ice (Gade, 1979). As ice melts, the source water doing the melting is cooled and freshened in a specific ratio depending on physical parameters such as the latent heat and specific heat capacities of water and ice. For instance, for the Ross Ice Shelf cavity, an analysis of the temperature–salinity diagram of the various water masses in the cavity, coupled with the melting–freezing (M–F) constraint mentioned above, has led to an improved quantification of the water mass transformations occurring there (Figure 16.7).

### FIGURE 16.7

A potential temperature–salinity diagram illustrating melting, freezing, and mixing processes that can occur beneath an ice shelf. Such diagrams can be useful in identifying the production of Ice Shelf Water (ISW) and thus inferring the amount of glacial basal melt into the ocean. By melting and freezing (M–F) at the base of the ice shelf, the properties of the ocean mixed layer at the base move only along the solid lines, which have potential temperature to salinity (Δθ/ΔS) slopes of 2.65. The “+” symbols represent observed water masses. The dashed lines indicate possible mixing between observed varieties of ISW and the high- and low-salinity shelf waters (HSSW and LSSW). This diagram is for the Ross Ice Shelf cavity. From Smethie and Jacobs (2005).

#### 3.6. Sea-Level Change

On the timescale of the earth’s glacial cycles, the volume of the global ocean changes substantially, being approximately 100 m or more lower in global sea level during a glacial period than the sea level of the present interglacial (Siddall et al., 2003; Chapter 27). This glacial cycle change in global sea level is thought to be forced by the amount of solar radiation reaching the earth, the periodicity of which is known as the Milankovitch cycles, which are dependent upon the precession, obliquity, and eccentricity of the earth–sun system (Imbrie et al., 1992). Remarkably, in these regular cycles of continental ice-sheet growth and decay, there are periods in which the sea-level changes abruptly. A remarkable example is the event known as meltwater-pulse 1A that occurred approximately 14,000 years ago during the last deglaciation (Stanford et al., 2006). Over a period of a few hundred years, the global sea level at that time rose by at least 5 m. The implication from this is that future global sea level could also change as rapidly.

The mechanism by which such rapid past change could have occurred remains obscure. One theory is that sudden import of warm ocean waters to the periphery of the ice shelves, particularly the marine-based portions, leads to a rapid wasting of the ice that triggers major change over a short time period. The change in ocean circulation would have to be driven by changes in atmospheric circulation, and the mechanism for that change is equally obscure. Continued collection of observational data and development of coupled global models will likely unravel this mystery in the coming years.
4. MARINE PERMAFROST

In the previous two sections (Section 2 and 3), we have discussed the interaction of ice that floats on the surface of the ocean with the upper ocean. Ice is also found on the bottom of the ocean, embedded in the sediment. There are two types of such ice, one being a pure ice matrix and the other involving trapped methane.

4.1. Pure Ice

During glacial periods, large portions of the Siberian Arctic continental shelves have been covered by ice, and the subsequent flooding of these shelves by ocean waters in interglacial times has resulted in large areas of what was originally land permafrost being transferred to the marine permafrost category. In general, marine (or subsea) permafrost is formed either by such inundations or by areas of seafloor with mean annual ocean temperatures below 0 °C. A more technical definition of marine permafrost calls for the seafloor ice structure to exist for at least a 2-year period (PAGE21, 2013), a condition easily met along the Siberian Arctic shelves. The thickness of the marine permafrost layer has been assessed using direct coring and reaches to 100 m in some areas. Electromagnetic techniques have also been employed based on the fact that sediments in an ice-bonded state have a different electrical resistivity than in the unbonded state (Corwin, 1983). Seafloor permafrost has been found only in the Arctic.

4.2. Methane Clathrates

There is a peculiar water-molecule-based structure, formed under modest pressure (depths greater than 300 m of ocean) and low temperature (below 0 °C), that results in the trapping of a gas such as methane in a cage surrounded by water molecules (Max, 2003). The solid is similar to ice (Figure 16.8a), except for the trapped foreign molecule. These clathrate hydrates collectively form a component of the marine permafrost layer. The clathrate can exist at depths in the marine sediments where the sediment is colder than the freezing point of the clathrate, itself a function of pressure (Figure 16.8b). Clathrates are suspected to be located along the Arctic coastline beneath the marine permafrost that has formed as a result of glacial advances, mentioned above. However, they are not restricted to such locations. Oceanic deposits seem to be widespread on the continental shelf and can occur at depth within the sediments or close to the sediment-water interface. They may cap even larger deposits of gaseous methane (Kvenvolden, 1995). It has been conservatively estimated that the amount of carbon in the clathrate marine form is more than twice that found in all fossil fuels on the earth (USGS, 2012).

FIGURE 16.8 (a) A methane clathrate block found in a depth of about 1200 m of water in the upper meter of the sediment. (b) Schematic profile of temperature through the ocean and uppermost marine sediments. The temperature for clathrate stability, \( T_s(P) \), increases with pressure (or depth). Experimental data for \( T_s(P) \) in pure water (dashed line) and seawater (solid line) are shown. The base of the hydrate stability zone (HSZ) is defined by the intersection of the Geotherm with \( T_s(P) \). Warming the ocean temperature, and hence the Geotherm, will shallow the HSZ. Panel (a) from http://en.wikipedia.org/wiki/File:Gashydrat_mit_Struktur.jpg; panel (b) from Buffet and Archer (2004).
5. EMERGING CAPABILITIES

In this section, we discuss the emerging capabilities in observing sea ice and the ocean beneath sea ice and ice shelves, some of the most difficult parts of the ocean to study. Improvements and future directions in sea-ice and land-ice modeling are also given; advances in ocean modeling are provided elsewhere in Chapter 20.

5.1. Ice-Capable Observations

Starting in the 1970s, the observation of sea ice was revolutionized by satellite remote sensing using passive microwave instrumentation, leading to unprecedented sea-ice concentration detail (Gloersen et al., 1992). Subsequent developments have led to sea-ice velocity fields determined from visual or active radar imagery (Emery et al., 1997). Now emerging is the possibility of determining the sea-ice thickness field, the most difficult of the sea-ice-related basic fields to measure by remote sensing (Laxon et al., 2003). The relevant instruments are the CryoSat radar altimeter (Wingham et al., 2006) and the ICESat laser altimeter (Schutz et al., 2005), which are able to give relatively precise observations of sea-ice thickness, and thus thickness change. This new ability to measure thickness is a key missing piece needed to more fully observe future Arctic sea-ice change, as it now allows the possibility, combined with the sea-ice concentration observations and sea-ice velocity, to make estimates of sea-ice volume and mass transport in the coming years.

A number of ice shelves in West Antarctica are undergoing significant mass loss due to what is assumed to be increasing rates of basal melt (Shepherd et al., 2004). There has been some efforts to determine basal melt rates using a combination of remote sensing data sets, basically surface elevation and surface velocity to infer ice flux divergence, and thus to evaluate basal melt (Joughin and Padman, 2003). As new satellite instruments come on board over the next few years, it is expected that observations of melt rate beneath all ice shelves will improve. However, these methods do have difficulties when the ice shelf is not in a steady-state configuration. Satellite altimetric methods can give an estimate of surface elevation change, which can then be used for corrections, but these in turn rely on snow compaction models.

Another important process of ice-shelf mass loss is that of calving. This is a process that can be well observed from space, given the number of satellites now in orbit producing high-resolution imagery (Herried et al., 2011). It will be possible to form an inventory of icebergs, starting from the birth to their ultimate demise. Such iceberg tracking will also provide a data set of freshwater melt input to the ocean from the icebergs in both polar seas (Gladstone and Bigg, 2002).

Aside from just taking an inventory of icebergs, there is still the nagging question of the physical processes that lead to calving in the first place (Bassis, 2011). To gain insight, it will be necessary to make measurements of both the strain and stress fields, in high spatial and high temporal resolution, dictating the need to make such observations in situ, either on the ice surface or with rapid, repeat airborne surveys. On the ice, the placement of newly developed passive seismic instrumentation may turn out to be useful to map out the stress field before, during, and after a major calving event (MacAyeal et al., 2009).

5.2. Ocean-Capable Observations

There are a significant number of emerging technologies to tackle the notoriously difficult problem of measuring ocean properties either beneath sea ice or in a sub-ice-shelf cavity. The subsurface polar ocean is one area where remote sensing by satellite has not been able to contribute much as the ice cover provides a surface obstacle to sensing. This implies that in situ observations are needed throughout the polar oceans. Perhaps the major breakthrough in this area has been the successful development and deployment of so-called ITP (Ice-Tethered-Profilers) (Krishfield et al., 2008; ITP, 2012). These instruments are profiling CTD devices that travel up and down a cable passed through access hole drilled through either sea ice (see Figure 16.9a) or an ice shelf. The data are retrieved via satellite in almost real time and have enormously increased the observational hydrographic data set in the Arctic basin (Toole et al., 2011). While thebulk have been deployed in the Arctic basin to date, their usage is now spilling over into the Antarctic. The Antarctic, because of its more seasonal sea-ice cover, is a more challenging environment for ITP deployment.

These ITP instruments are in some ways analogous to the Argo float program. Argo floats, of course, drift freely with ocean currents and surface from time to time to transmit their data (Send et al., 2009), but obviously this strategy does not work well, or at all, in an ice-covered environment (Klatt et al., 2007). Traditional Argo floats are now being ice hardened, so they can, if the opportunity arises, pop up through leads in sea ice (see Figure 16.9b). Another issue is that Argo floats determine their position by reaching the ocean surface and using a GPS receiver but, again, cannot do so in an ice-covered environment. If they are to operate in a region where they cannot surface often enough to satisfactorily fix the position of the data they are measuring, they need to use acoustics to navigate (Howe and Miller, 2004): arrays of subsurface acoustic beacons are currently being deployed to provide the necessary acoustic infrastructure.

A shortcoming of both the freely floating ITP and Argo buoy is that they are not able to be navigated to specific locations, for example, to perform repeat track surveys.
The use of AUV (Autonomous Underwater Vehicles) has been successful in exploring beneath sea-ice cover and ice shelves for short periods of time (McPhail et al., 2009). Such vehicles are now being engineered to have long range and long duration, including the capability to “hibernate” on the seafloor for extended periods of time. Although this is a hazardous enterprise, steps can be taken to minimize the risk (Jalving et al., 2008). For more sustained missions, lasting several months, and possibly up to a year in the near future, the development of the

FIGURE 16.9 Dominant, emerging observational technologies used to observe the polar ocean beneath the sea-ice and ice-shelf covers. (a) Ice Tethered Profiler (ITP) anchored to ice, (b) Argo float modified for under-ice deployment,
underwater glider (see Figure 16.9c) perhaps offers the greatest hope of obtaining hydrographic sections in ice-covered oceans (Eichhorn, 2009). Like the Argo buoy for under-ice missions, the gliders do need navigational assistance from subsurface sound beacons. As an example, such a glider setup has been recently deployed for Davis Strait in Baffin Bay (Lee et al., 2004).

The ubiquitous presence of marine mammals in the polar oceans, such as seals, has led to the development of the SRDL (Satellite Relay Data Logger). This is a miniature CTD, coupled with a small Argos transmitter, that is attached to the back of the neck of a seal (Fedak et al., 2002). The seal is captured in a net, held for approximately 15 min during which a fast-acting glue, attaches the SRDL to the seal (see Figure 16.9d) (Teilmann et al., 2000). Such data sets have the limitation that they only provide data for approximately 1 year, as the SRDL falls off the seal during the annual molt. Another issue is that the data are only collected wherever the seal decides to travel. However, as the behavioral patterns of various seals species is now beginning to be learned, the SRDL are becoming more useful for data collection in selected regions of the polar oceans (Campagna et al., 1999).

A surprising property of a fiber-optical cable is that laser light shot along the length of the cable can be used to determine the temperature at approximately every 1 m for more than a 10-km stretch (Bao et al., 1993). This is possible because the scattered light can be analyzed by a detector, and based on physical principles of scattering, along with in-field calibration of the cable, one can obtain a time series of temperature along the cable at frequent time and space intervals (Selker et al., 2006). This technique is known as DTS (Distributed Temperature Sensing) (see Figure 16.9e for a typical deployment). Such a cable was recently deployed through the Ross Ice Shelf and a year-long data set of ice-shelf and sub-ice-shelf temperatures was collected (Stern et al., 2013).

5.3. Ice-Capable Modeling

Traditionally, sea-ice models have been developed based on finite difference discretizations of the governing equations and solved on either a regular Cartesian or a latitude–longitude grid (Holland et al., 1993). There are, of course, much more advanced discretizations and gridding schemes now developed in other areas of science. Ice researchers have begun to apply such techniques, for example, finite element modeling (Timmermann et al., 2009) which allows for very irregular grids (see Figure 16.10a). Such grids can be refined to very high resolution in narrow passageways and channels such as in the Canadian Arctic Archipelago (Liettaer et al., 2008).

While discretization and gridding dictate the spatial accuracy of the numerical solutions, time-stepping efficiency is also a major issue. Global-scale climate models seek to make each component submodel as efficient as possible. For the most part, sea-ice models have used explicit time-integration schemes (CICE, 2012), but more recent efforts have sought to bring implicit, matrix-based, methods to bear on the problem. For instance, a matrix-free Newton–Krylov technique (Lemieux et al., 2012) has been applied to the sea-ice equations and shows promise for allowing a significant speed up in integration time.

Numerics aside, aspects of the basic physics of sea ice still need to be improved. Parameterizations of a host of subgrid-scale phenomena continue to be tackled by researchers; for instance, summer melt ponds and their drainage, brine inclusions, and surface albedo (Holland et al., 2001). On the dynamics side, there is still uncertainty regarding the underlying rheology with the earliest rheology focusing on isotropic behavior and more recent studies looking into the anisotropic behavior of sea ice (Feltham, 2008). The most common rheology currently in use is a combination of elastic, viscous, and plastic behaviors (Hunke and Dukowicz, 1997). One feature still missing in all these rheologies is that of a tensile strength, appropriate to the description of landfast ice, as seen along parts of the coast of the Arctic and Antarctic (König Beatty and Holland, 2010). Current rheologies have not incorporated this phenomenon, but it is needed as the presence of landfast ice radically alters the formation of coastal waters, feeding the deep ocean (Barber and Massom, 2007).

The modeling of land ice, including that of ice shelves, has lagged significantly behind that of sea ice, but recent efforts, bolstered by the need for land-ice modeling in IPCC class models, are beginning to change that situation (CISM, 2012; see Figure 16.10b). The task is significant as it requires major advances in both the numerical modeling and the understanding of the physics of land ice and ice shelves, and also in the coupling of the ice models to ocean general circulation models. To date, there has not been a plausible coupling in which the ocean volume and ice volume swap physical space with one another when either the ocean melts the ice and invades its physical space or vice versa (Lipscomb et al., 2009).

6. CRYOSPHERIC CHANGE

In this final section, we look back on recent changes in the cryosphere and look ahead to projections for sea ice, land ice, and marine permafrost in the present century and beyond.

6.1. Observed Sea-Ice Change

The radical reduction in Arctic sea-ice areal extent over the past few decades has caught the world by surprise. It was not so long ago, in the early 1990s, when the Arctic sea ice was thought to be a stable, robust feature of the earth’s
FIGURE 16.10 Examples of emerging technologies in sea-ice and ice-sheet (shelf) modeling. (a) Sea-ice-thickness modeling based on highly variable grid resolution capable of resolving flow through narrow straits. (b) High-resolution model of the flow of the entire Greenland Ice Sheet. Note the bright red areas showing the speedup of the ice sheet as it emerges toward the ocean through narrow outlet fjords. https://www.uclouvain.be/en-376557.html; http://public.lanl.gov/sprice/images/GIS_vels.jpg.
modern climate system, at least on the short timescale of years to decades. However, satellite remote sensing has unequivocally shown a nearly 50% decrease in the extent of the Arctic pack ice during the summer minimum (Figure 16.11, blue curve). In fact, the past year (2012) has set an all-time record low (NSIDC, 2012). The cause of this decline is tied up with the occurrence of global warming over the past century or so (Stroeve et al., 2007), as the ice-albedo feedback effect, once it takes hold, contributes to the accelerated melting. There is of course natural variability in the Arctic sea-ice cover, as driven, for example, by the Arctic Oscillation, a pattern of winds that can either reinforce ice export out of the Arctic or hold it within the basin (NOAA, 2012). Coupled climate models have been successful in hindcasting Arctic sea-ice decline in general, but none of the models to date have been able to accurately hindcast the actual extreme decline, nor the extreme minima in past years, 2007 and 2012 (Rampal et al., 2011). This indicates that the coupled models are not yet fully able to capture extreme events, suggesting more research is needed either in the individual model components or in the model couplings to advance this capability.

The story with the Antarctic sea ice is just the opposite, with a possible small trend upward in extent (Figure 16.11, red curve), perplexing sea-ice researchers, and global climate modelers. While it is thought that the ice-albedo is playing a major role for Arctic sea ice, it is confounding that no such mechanism appears to be at play in the Antarctic. Of course, a major distinction between the Arctic and Antarctic is that in the former the sea ice is present in the summer in appreciable quantity but the Antarctic sea ice largely disappears, thus possibly reducing its susceptibility to the ice-albedo feedback effect. Additionally, the Arctic is effectively landlocked by the surrounding continents, whereas in the Antarctic, the inverse is the case and this may play a role. There is some evidence that wind patterns in the Antarctic have controlled the Antarctic sea ice (Holland and Kwok, 2012). Still, the apparent lack of an ice-albedo feedback remains a major question for understanding global climate change and the cryosphere.

6.2. Sea-Ice Projections

On the short timescale, looking ahead just a few months, an international initiative called the “Sea Ice Outlook” (SEARCH, 2012) has joined together a large number of sea-ice modeling efforts so as to provide a summary of the expected September Arctic sea-ice minimum. Monthly reports are released through the summer in an attempt to provide the scientific community and stakeholders with the best available information.

On the longer decadal and centennial timescales, sea-ice hindcasts and projections have been organized under the IPCC rubric, which has put together simulations based on ensemble runs of those global climate models that contain some form of a sea-ice component. The simulations captured to some degree the climatological annual mean, seasonal cycle, and temporal trends of sea-ice area during the modern satellite era, although there was considerable


Arctic (blue curve) and Antarctic (red curve) standardized anomalies and trends of sea-ice cover over the period 1978–2012. Thick lines indicate 12-month running means, and thin lines indicate monthly anomalies. The Arctic shows a definite trend downward, while the Antarctic shows a slight increase.
scatter among the models. Each member of the ensemble performed differently, and the interannual variability within each member model is not yet fully understood. An interesting aspect of the simulations is that multimeter ensemble means show promising estimates close to observations for the late twentieth century (Zhang and Walsh, 2006). With regard to sea-ice thickness distributions, considerable deficiencies in sea-ice properties in many model results have been seen. Most of the models have difficulty in reproducing the spatial sea-ice thickness distribution of the past, particularly in summer (Gerdes and Köberle, 2007). Taken collectively, the models indicate a significant reduction in Arctic sea-ice cover over the coming decades. However, the large scatter between individual model simulations at present leads to much uncertainty as to when a seasonally ice-free Arctic Ocean might be realized (Serreze et al., 2007). Of real concern is the fact that the models failed to capture the extreme minimum that occurred in 2007 (prior to the issuance of the IPCC report). An important criterion for future development and improvement of sea-ice models will be their ability to capture such extreme events.

Projections from IPCC Antarctic sea-ice models show there is larger uncertainty in model performance in the Southern Hemisphere (Arzel et al., 2006) with a much larger disagreement between the model ensemble members for the Southern Ocean sea-ice simulations. Many of the models have an annual sea-ice extent cycle that differs markedly from that observed over the past 30 years (Turner et al., 2013). In contrast to the satellite data, which exhibit a slight increase in extent, the mean extent of the models over the period 1979–2005 shows a decrease (Turner et al., 2013). Several studies have shown that the response of the Southern Ocean to an increase in atmospheric greenhouse gas concentrations is delayed compared with other regions (Goosse and Renssen, 2005), mainly because of the large thermal inertia of the Southern Ocean. This highlights the earlier discussion of the bathymetric contrasts between the Arctic and Southern oceans, the fact that the former is largely disconnected from the World Ocean while the latter is not. Aside from an inability to properly capture the recent ice extent trends, the coupled models have also not been able to capture the regional spatial variability of sea-ice cover over the Southern Hemisphere in the past few decades, during which time the Ross Sea has been gaining cover while the Amundsen has been losing it (Stammerjohn et al., 2008). Again, this suggests need for further research to improve sea-ice models, and their coupling with the ocean and atmosphere.

### 6.3. Observed Land-Ice Change

The recent breakup of ice shelves in both Greenland (e.g., Jakobshavn) and Antarctica (e.g., Larsen B) has also caught researchers by surprise. The Jakobshavn Ice Shelf has been undergoing periodic retreat over the past 100 years or so (Sohn et al., 1998), but with a dramatic thinning and retreat occurring starting in 1997 (Joughin et al., 2004). No less dramatic was the breakup of Larsen B, occurring in just a few weeks in early 2002. From an analysis of the marine sediments located beneath the former Larsen B Ice Shelf, it has been demonstrated that the recent collapse has been unprecedented in the Holocene (Domack et al., 2005). There is some evidence to suggest that the Greenland breakup was forced by the arrival of warm waters (Holland et al., 2008) and that the Antarctic case was driven by the arrival of warm air over the Antarctic Peninsula (Marshall et al., 2006). In the case of both Jakobshavn and the Larsen B ice shelves, the inland glaciers connected to the shelves have sped up dramatically, with a doubling or betterment in their speeds (Rignot et al., 2004). The breakup of these ice shelves has led to a refocus toward both the sensitivity of these ice shelves to climate change and the implications of ice-shelf decay for the stability of grounded ice (De Angelis and Skvarca, 2003).

A far more disconcerting example of recent glacial change is the observed thinning of several of the ice shelves fringing the Amundsen Sea in West Antarctica (Pritchard et al., 2012). As those ice shelves serve as a front to the massive marine-based West Antarctic Ice Sheet, any negative mass balance signal in that area could hold serious consequences for global sea-level rise. Using satellite laser altimetry combined with modeling of the firn layer, it has been deduced that ice shelves in this sector of Antarctica are thinning through increased basal melting (Pritchard et al., 2012). In fact, the greatest thinning occurs where warm, deep water can gain access to the continental shelf, and ultimately the grounding zone of the glaciers, via deep troughs in the shelves. From ocean modeling studies, it appears that slowly changing wind stress patterns over the Amundsen Sea are responsible for the changes in warm water volume being brought aboard the continental shelf (Thoma et al., 2008). Understanding what controls the variability of the large-scale wind patterns in the polar regions and indeed on the global scale is an area of active research and holds the ultimate key to how warm water can arrive at the base of an ice shelf (Kerr, 2008). Both Greenland and Antarctica are showing an overall accelerating mass loss (Figure 16.12).

### 6.4. Land-Ice Projections

The projections of land ice change, and the implicit role that climate change and warm ocean waters could have on breaking up of ice shelves and of causing acceleration of inland ice into the ocean, have been lacking. This shortcoming is implicitly acknowledged by IPCC which failed to provide an estimate of the potential dynamical contribution of glaciers to sea-level rise over the present century.
This apparent shortcoming serves to underscore the difficulty of this task, which the research community has only relatively recently addressed in earnest (IPCC, 2010). The projections of the contribution to sea-level rise are uncertain (Alley et al., 2005) as they range from a few centimeters rise over the next century to a meter or more, the latter implying devastating consequences for coastal habitats and infrastructure (Nicholls and Cazenave, 2010). Part of the difficulty in producing credible projections is the lack of observational evidence upon which to base such model (Alley et al., 2005). That situation is changing with the emerging ice and ocean observational technology as discussed in Section 5, but it will take decades to build up meaningful observational records in these difficult-to-access regions. Progress in model coupling is likely to proceed at a faster pace (Lipscomb et al., 2009), but will have to await the availability of basic observation data, again highlighting the need for the science community to have a sustainable ocean and ice observational system in place around the periphery of the great ice sheets.

6.5. Marine Permafrost

Concern has been raised that the considerable deposits of methane, in the form of sub-seafloor clathrates or gas trapped beneath other permafrost, could be released under a scenario of global warming leading to a runaway feedback on global warming. Methane is a powerful greenhouse gas, having a global warming potential of more than 60 times that of carbon dioxide. Over the Arctic continental shelf, methane gas release has indeed been detected (Kort et al., 2012). The observed distribution of dissolved methane and possible mechanisms of gas release in connection with observed dynamics of coastal ocean environments suggest that areas of the Arctic shelves are important natural sources of methane to the atmosphere and that such areas tend to be affected by ongoing global change (Shakhova et al., 2007), possibly being susceptible to ongoing and future changes in sea-ice and ocean circulation.

The speculative scenario of a release of vast amounts of methane or its clathrate form is commonly referred to as the clathrate gun hypothesis (Kennett et al., 2003), the main idea being that the seafloor permafrost would itself initially melt under conditions of a warming global ocean thereby triggering the release of methane. In its original form, the hypothesis proposed that the clathrate gun could cause abrupt runaway warming on a timescale of less than a human lifetime (Kennett et al., 2003) and might have been responsible for warming events in and at the end of the last ice age (Behl, 2000), but more recent research now suggests this to be unlikely (Sowers, 2006). Past suspected releases during the Permian–Triassic extinction event and the Paleocene–Eocene Thermal Maximum have been linked to such gas release and past climate change (Benton and Twitchett, 2003).

7. SUMMARY

In this chapter, we have looked into aspects of the interaction of the World Ocean with the marine cryosphere, focusing on the Arctic and Antarctic sea-ice covers, the outlet glaciers of the Greenland and Antarctic ice sheets, and to a lesser extent, the marine permafrost beneath the seafloor. We have discussed the fundamental chemical and physical properties of freezing water and ice that organize the manner in which ice forms or melts in the ocean, be it sea ice or land ice, and have seen that heat exchange at the ice–ocean interface is unique in ocean dynamics because of the fundamental role played by molecular exchange. The description of these fine-scale processes and their encapsulation as parameterizations for use in global climate models is a maturing subject.

On the large scale, the transport of heat to the polar ocean is also relatively well understood, but what is not understood is the manner in which that heat transport changes through natural variability or anthropogenic forcing. The global ocean impacts the marine cryosphere by transporting warm water to the base of sea ice or the base of an ice shelf, and the ensuing melt impacts back onto the global ocean through the conveyor belt by both influencing surface convection and bottom water formation. The ramifications of these large-scale ice–ocean interactions and potential feedbacks on the climate system are still an area of active research.

Remote-sensing observational technologies have been delivering data sets that provide an unprecedented global
view into the marine cryosphere. Perhaps the largest advance, and surprise, of the past decades has been the witnessing of the enormous change occurring in both sea ice and land ice. This change has largely reshaped a viewpoint in which the marine cryosphere was thought to be a relatively fixed feature of the climate system on the timescale of decades or a century. We have now begun to see that viewpoint unravel and be replaced by one where the new norm is Arctic sea ice being halved in summer extent and a number of outlet glaciers doubling their seaward velocity. Such change shows the sensitivity of the marine cryosphere to climate change, whether natural or anthropogenic, and the pressing need for further sub-ice, polar ocean observational data sets and numerical modeling to provide a credible and robust ability to project the future of the marine cryosphere.

REFERENCES


Nansen, F., 1902. The Oceanography of the North Polar Basin. Longmans, Green, and Company.


Ross, J.C., 1847. A Voyage of Discovery and Research in the Southern and Antarctic Regions V1: During the Years 1839–43. Albemarle Street, London.


