Chapter 6
Simulating the Ocean and Sea Ice

The ocean covers about 70 % of the surface of the earth. Water is more dense than air: The top 33 ft (10 m) of ocean has the same mass as that of the entire atmosphere. As long as we remain near the surface of the earth (in the climate system), mass is equivalent to weight. In addition, the heat capacity of water (for the same mass) is four times larger than air. Thus, it takes more energy to raise the temperature of the ocean by the same amount for the top 33 ft (10 m) than for the entire atmosphere above it. Putting it another way: The top 10 m of ocean holds more energy than the atmosphere above it. Including the rest of the ocean below 10 m, the ocean is a much larger reservoir of heat than the atmosphere. Thus, the heat content of the ocean is a critical part of the climate system.

The ocean is also "stratified" with a series of shallow surface circulations and a deep ocean circulation (see Chap. 2). Large parts of the deep ocean do not interact rapidly with the ocean surface and hence can store heat away from the atmosphere. The ocean can serve as both a source and a **sink** of heat to the surface climate system (the atmosphere and land) on very different timescales from days to weeks up to hundreds of years. Thus, the ocean acts like a giant and slow reservoir that holds and redistributes energy in the climate system. It can also store carbon in several different forms, and that also makes it an important part of the carbon cycling through the climate system.

Simulating the ocean is critical for understanding climate on many scales. Some of the critical aspects are the exchange of heat and water with the surface, and the role of heat and **salinity** (the proportion of salt) in altering the density of the ocean. Density plays an important role in the ocean: Heavy water sinks; light water rises. The general nature of the **ocean circulation** cannot be understood without it. The basic elements of the ocean circulation are a result of the ocean boundaries (topography), the rotation of the earth, surface winds, and the changes to water density by changing the heat and salt content of water. These factors ultimately drive the ocean circulation, and they need to be represented in ocean models and properly coupled to the other parts of the climate system.

The cryosphere ("ice" sphere) contains land ice (ice sheets and glaciers), seasonal snow on land, and sea ice. Because land ice and snow are linked to land models, we discuss them in Chap. 7, on terrestrial systems. But it is logical to discuss models of sea ice in this chapter, as they are tightly coupled to the ocean. Sea ice is a critical

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part of the climate system because it strongly affects the surface albedo (white ice is much brighter than the dark, ice-free ocean, and it reflects more light back to space), and it also affects the surface energy coupling between the atmosphere and ocean (acting like an insulating blanket that allows the ocean to retain heat). Because of this, even though the cryosphere is a small area of the planet, it is an important part of the climate system, and it is critical at high latitudes.

In this chapter, we discuss ocean models and compare and contrast them with atmosphere models. We also cover models that simulate sea ice.

6.1 Understanding the Ocean

The key aspects of the ocean in the climate system can be described by understanding the ocean structure, what drives or forces the ocean, and how this gives rise to the ocean circulation.[1] Modeling the ocean requires a representation of its structure. As with the atmosphere, the critical part is understanding what processes force the ocean, and then creating an appropriate representation of the physics of the forcing and structure that give rise to the ocean circulation.

6.1.1 Structure of the Ocean

Figure 6.1 is a simple schematic of the ocean. Think of it as a cross-section of one ocean basin, like the Atlantic from the South Pole to the North Pole. The ocean is divided vertically into a **mixed layer** near the surface (the top 50–100 m), where the ocean interacts rapidly with the atmosphere. Below the mixed-layer region, the effect of mixing with the surface gets smaller. Then deeper in the ocean (usually several hundred meters below the mixed layer) is a region of sharp gradients called the **thermocline** (*thermo* = heat and *cline* = gradient). Properly, this gradient is the **pycnocline** (*pycno* = density). The **surface ocean** lies above the pycnocline and contains the mixed layer. Below the pycnocline gradient, water is much colder and denser, and it exchanges very slowly with the surface. This is the **deep ocean** (or "abyss"), where the water temperature is nearly uniformly cold (pretty close to freezing). In high latitudes near the poles, the water is nearly the same temperature from the surface to the bottom, and there is usually not a thermally driven density gradient separating the surface and deep ocean. The density gradient in high latitudes is provided by changes in salinity (**halocline**; *halo* = salt).

The ocean in the tropics and mid-latitudes is stratified by density, with lighter water (usually warmer) on top and cold water beneath, illustrated by warm and cool

[1] A good detailed but qualitative treatment for the general reader is in Vallis, G. (2012). *Climate and the Oceans*. Princeton, NJ: Princeton University Press.

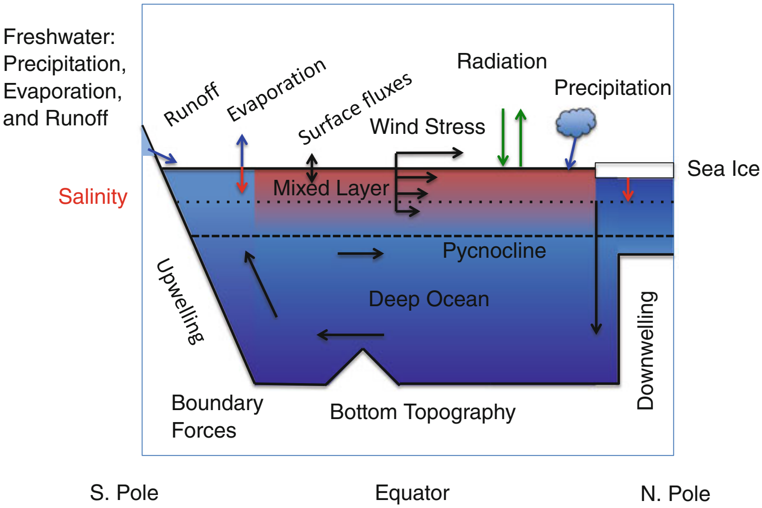

Fig. 6.1 Ocean schematic. This view of the ocean shows the mixed-layer boundary (*dotted line*) and the pycnocline or thermocline boundary (*dashed line*), with the deep ocean below. Changes of salinity are in *red*; changes in water are in *blue*; and evaporation, precipitation and runoff, and changes in energy (radiation) are in *green*. Ocean motions are in *black* and include wind stress and surface currents, and downwelling and upwelling in the deep ocean

colors in Fig. 6.1. At high latitudes, the temperature is cold all the way from the surface to the bottom (all cool colors). This means that, at high latitudes, surface water and deep water can easily mix if the density of one changes slightly (less dense water rises; more dense water sinks).

This structure of a mixed-layer, thermocline, and deep ocean arises from the interaction of the rest of the climate system with the ocean, and the properties of salty water. The density of water is a critical part of understanding the ocean. Ice is less dense than liquid, and it floats. In the same way, fresh water is less dense than salty water (salinity increases density), and warm water is less dense than cold water. Thus, when ocean water cools or acquires more salt, it gets more dense and, if denser than the water below it, sinks.

6.1.2 *Forcing of the Ocean*

The mixed layer is forced by **surface fluxes** (exchanges with the atmosphere). These exchanges include both exchanges of heat, momentum, and masses of salt and water. Heat enters the ocean at the surface by solar radiation filtering through the atmosphere. The exchange of heat, and the evaporation of water, changes ocean temperature. Changing temperature also changes the density of water. There are exchanges of freshwater between the ocean and atmosphere by precipitation and evaporation, and even from the land surface through rivers (see Fig. 6.1). Evaporation requires energy and removes heat from the ocean. Freshwater (without salt) also leaves the ocean when sea ice forms. Salt is expelled as sea ice forms. Freshwater returns to the ocean as sea ice melts (which might be a season or a year later). The addition or removal of water generally conserves salt content in the ocean, so as water is added or subtracted, the density changes. Atmospheric winds also push on the ocean, transferring momentum (often called "**stress**" on the ocean). The stress from wind on the surface generates surface ocean currents. These processes are indicated in Fig. 6.1. Note that almost all of these processes will affect the ocean circulation (cause water to move) either by pushing it (wind stress), or by changing heat or salinity that alters density (causing water to rise or sink).

Cold, dense surface waters at high latitudes form a link to the deep ocean, below the pycnocline. The deep ocean is driven by the density-driven motion of water, as well as by boundary forces on the side and bottom, and ocean topography (**baythemetry**) that alters circulations is similar to the way mountain ranges alter the atmospheric flow. Note that the ocean has boundaries on all sides and boundary layers on all sides. The atmosphere, however, has a boundary layer only at the bottom, with basically no mass at the top.

6.2 "Limited" Ocean Models

Like the atmosphere, simulating the ocean dynamically with a finite element model is a difficult task, requiring lots of approximations. As with the atmosphere, there are several types of ocean models. We concern ourselves mostly with ocean general circulation models (GCMs), which are similar to GCMs for the atmosphere. Ocean GCMs have complex representations of the ocean circulation throughout its depth, full ocean topography, and parameterizations for small-scale mixing processes at the surface and within the ocean.

Beyond the full representation of the ocean structure in Fig. 6.1 are simpler types of ocean models. These models try to represent individual ocean basins or just regions of those ocean basins (Fig. 6.2). As with models of the atmosphere, they are often called regional models. Basin-scale regional models may be used for experiments that look at coupling between the ocean and the atmosphere. Regional ocean models are similar to limited area atmosphere models discussed in Chap. 5: They are often used at high resolution, coupled to limited area atmosphere models to represent details of regional weather or climate. They are forced at their boundaries by observations or by output from other models. These regional models may not include a deep ocean circulation. The deep ocean circulation is often specified in regional ocean models. Regional ocean models are designed primarily to represent the region above the thermocline that changes relatively quickly and responds to the surface climate system. These models are focused on the surface properties of the ocean that interact on fast timescales (days to weeks or seasons) with the atmosphere, ice, and land surface. They have detailed representations of surface

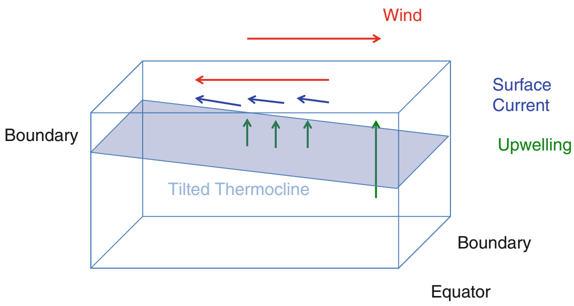

Fig. 6.2 Ocean basin. Representation of an ocean basin from the equator to a northern boundary. Trade winds near the equator (*red*) cause water to move along and away from the equator, also tilting the thermocline or pycnocline. The motion of water away from the equator causes upwelling along the equator

exchanges, radiation, and wind stress that forces the mixed layer of the upper ocean. They may have detailed representations of the ocean floor topography, often near coastlines.

A commonly used type of global ocean model seeks only to represent surface processes, and the fast communication of the ocean with the rest of the climate system. These are global **mixed-layer ocean models** (sometimes referred to as “slab” ocean models). Mixed-layer ocean models are global in scope, representing all ocean basins, but they are focused on the upper ocean, not the deep ocean, that is, the region above the pycnocline in the mixed layer in Fig. 6.1.

The advantage of mixed-layer ocean models for simulating the climate system is that they contain smaller amounts of water mass (from 10 to several hundred meters deep). Like the limited area models described earlier, these models have a limited area, but here the limit is in depth. The models typically have no circulation but are simply an energy balance equation: The temperature of a fixed-depth layer is determined by the heat and water coming in and out at the surface, and the specified heat from a bottom boundary. They do not include currents or water motions.

A mixed-layer ocean model just adjusts the temperature of the slab of water to respond to the surface energy budget from the atmosphere, and an assumed interaction with the deep ocean below. Mixed-layer ocean models need to have a bottom boundary condition, usually a specified movement of heat to and from the unresolved deep ocean below. Like a limited area model in the horizontal, the boundary conditions often come from observations or from a more detailed ocean GCM. The models are designed to reproduce the ocean surface temperature and interact with the rest of the climate system given a specified ocean circulation pattern below. The advantage of this configuration is that it reduces the heat capacity of the ocean and allows it to come into balance much faster: decades rather than centuries.

The disadvantage of mixed-layer ocean models is that they require a fixed assumption about the ocean circulation, so if the climate is too different from what is assumed by the specified energy transfer from the deep ocean, the ocean model may not yield a reasonable result. Because water cannot move horizontally in this class of models, if too much heat is removed or added to a grid box in the ocean, the bottom heat exchange is fixed and the temperature of the surface temperature may change a lot. In a full ocean model, however, a change in surface heat exchange may change density and cause water to move. The motion of water can take heat away from the surface and alter the heat exchange with the deep ocean, which is fixed in mixed-layer models. So mixed layer ocean models need to be used with some caution, and for more reliable climate system calculations, full models with the ocean circulation are used.

6.3 Ocean General Circulation Models

We divide our discussion of ocean modeling into a description of the grids and dynamics used in an ocean model, the deep ocean and the thermocline, and the surface ocean and the mixed layer. The surface ocean and the mixed layer are of primary concern to many of the simplified models described above as well.

6.3.1 Topography and Grids

As with the atmosphere, GCMs of the ocean are finite element models. However, whereas atmosphere models need be concerned only with a bottom boundary, and energy input at the top, the ocean has top, bottom, and side boundaries. The bottom boundary is similar to the topography we are familiar with on land: The ocean has mountains and valleys and complex topography that affects the circulation. The ocean has a top boundary that is critical for coupling to the bottom boundary of the atmosphere, and the primary forcing of the ocean occurs here. The horizontal boundaries of the ocean basins mean that the ocean grid is not global: Not every point needs to be represented, since some latitudes and longitudes have no ocean.

Note that these grids may change over time if the sea level changes. During the last ice age, when the sea level was about 330 ft (100 m) lower, there were significant differences in topography (bathymetry): The Bering Strait between Siberia and Alaska was a land bridge (enabling *Homo sapiens* to walk to the Americas from Asia), and the region between Indonesia and Australia was mostly land as well (with one or two channels).

Because the ocean does not occupy the whole planet, different grids have been constructed. First, a latitude-longitude grid is possible, but with boundaries and cells that do not exist because they are land points. However, having grids that converge into a single pole creates problems because the same number of cells exists at all longitudes. For a 1° longitude there are 360 points at the equator, and each point represents 68 miles (110 km). But at the poles, the distance around the earth goes to zero, so the size of the points becomes small. At 80° latitude, each degree of longitude is only 12 miles or 20 km. This introduces computational problems. Wind or currents may move air or water farther than one grid box in a single time step. For many processes, this creates problems in accounting for all the energy and mass. In the Southern Hemisphere, the pole and regions around it are conveniently on land (Antarctica) and the South Pole has no ocean, so this problem is not as acute. But the Arctic Ocean is difficult to represent on a traditional latitude-longitude grid.

Two approaches can get around this problem of having a pole in the ocean. "Equal area" grids are possible, where faces of a cube are projected onto the sphere so that most grid points have similar area. Another method is to "shift" the pole of an ocean grid onto land. The mathematics is complex, but this approach is computationally efficient and avoids mathematical problems with very small grid boxes (wedges) at the poles.

Another type of grid is not regular, and thus is often termed "unstructured." These **unstructured grids** often have finer resolution in critical areas for either global or regional simulation. The goal is to increase resolution where it matters, for example, in regions with narrow straits or important bottom topography, to better represent the boundaries of the ocean basins. This puts resolution where it is needed and is more efficient than having high resolution everywhere. Because ocean topography is such an important forcing term for ocean circulations (more so than in most regions of the atmosphere), these variable resolution grids are more common in ocean models.

6.3.2 Deep Ocean

The deep ocean (below the mixed layer) has a global circulation called the **meridional overturning circulation** (Fig. 6.3).[2] The circulation has one component driven by **buoyancy**, where water sinks because it gets heavier. Colder water is heavier and saltier water is heavier, so it convects when it sits on top of lighter water. **Convection** is the same buoyancy-driven force in the atmosphere, where lighter (warmer) air lies below heavier air, and it rises, forming clouds. The buoyancy component of the ocean overturning circulation is called the **thermohaline circulation**, driven by heat (*thermo*-) and salt (*-haline*). Surface winds also help drive the circulation. The ocean circulation is regulated by ocean topography that controls where water flows. Both processes interact to produce the deep ocean circulation: The density-driven thermohaline circulation acts like a "heat engine," whereas the components driven by surface winds might be more analogous to a "pump." It is the combination of these forces that results in the deep ocean circulation.

[2]Figure 6.3 is based on a figure from Rahmstorf, S. (2002). "Ocean Circulation and Climate During the Past 120,000 Years." *Nature*, *419*(6903): 207–214. doi:10.1038/nature01090. This paper is also a good introduction to the ocean circulation.

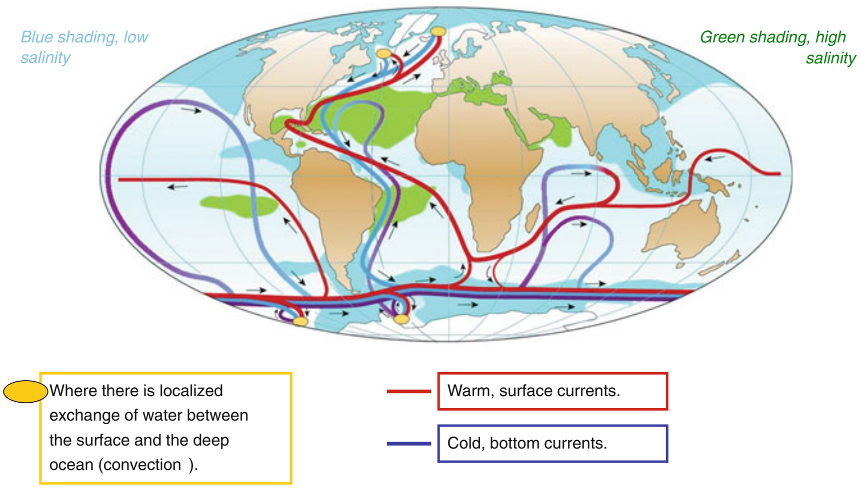

Fig. 6.3 Deep ocean circulation. The schematic shows the warm surface currents and deep bottom currents. *Orange* regions indicate where deep water forms. *Blue shading* indicates low salinity, and *green shading*, high salinity. Deep water forms in the North Atlantic and near Antarctica and flows throughout the ocean basins. Figure adapted from Rahmstorf (2002)

Representation of the ocean circulation in models is dependent on a number of factors. Of prime importance are the processes regulating the changes in density driven by salinity changes and temperature changes. In polar regions, sea ice forms (expelling salt into the ocean) and the ocean cools by losing heat to the relatively colder atmosphere. Both cause increases in density, and the surface water becomes denser than the water beneath. This water sinks to the bottom of the ocean, forming **bottom water**. Bottom water formation happens in both the Arctic and the Antarctic in the orange regions in Fig. 6.3. Antarctic bottom water forms just off the Antarctic ice shelf and is generally denser due to very cold temperatures, flowing throughout the bottom of the earth's oceans, constrained by bottom topography (see Fig. 6.3). Because masses of water do not mix very well, the ocean water keeps its properties (heat, salt, and trace chemical distribution) for long periods of time. Water is often named for, or characterized by, where it last encountered the surface. The oceans are cold at the bottom because the water comes from high latitudes. There is little warmth from the seabed. Water is warm when it is warmed at the surface by the sun.

The bottom water flows equatorward in the Atlantic (south). In the Pacific and Indian oceans it flows equatorward from Antarctica (north), starting a global circulation (see Fig. 6.3). The newly formed bottom or deep water may not see the surface again for a thousand years, and it will largely preserve its characteristics of salt and heat throughout that time, changing only slowly. Ocean circulation speed can be observed by looking at chemicals[3] in deep ocean water: Long-lived and inert industrial chemicals from human origins in the air mix into seawater in trace amounts and can be seen spreading out with the deep water in ocean basins, such as from the North Atlantic into the tropical Atlantic. The chemicals are present in microscopic quantities, and are inert and not dangerous. The presence of these industrial compounds provides a record of when the water last was at the surface.

One of the key processes to correctly represent the buoyancy-driven circulation is the vertical mixing of water masses as the denser (colder, saltier) water sinks. This is a density-driven convective process, physically the same type of process that occurs during atmospheric convective motions that drive deep thunderstorm clouds. Simulating the mixing that occurs during convection is important for representing the resulting composition (heat, salt) that the bottom water has. Typically the process is represented either as a simple adjustment necessary to get the water column "stable" again, with heavier water on the bottom, or as a "diffusion" process that depends on the large-scale vertical density gradient. Ocean vertical mixing is a critical parameterization in which the representation of a fundamental physical process (density) must describe complex interactions in space and time that are often below the resolution of the model.

In general, ocean models can do a good job of simulating the thermohaline circulation patterns, which are governed by the large-scale position of the continents. However, the amount of deep water that forms and, hence, the "speed" or mass in the overall circulation can vary quite a bit. The overall mass transport and speed of the circulation is dependent on a balance of processes in an ocean model: formation of deep water by density changes in the Arctic and the Antarctic, wind-driven upwelling around Antarctica, bottom topography and forces throughout the ocean, and diffusion of heat in the interior of the ocean. Surface stress from wind alters surface currents. The surface stress of winds around Antarctica pushes water offshore, and creates cold conditions where sea ice forms, also creating cold and salty dense water that forms the Antarctic bottom water. Finally, the surface return flows, such as the Gulf Stream (see below), transport their mass in eddies and a mean circulation. The flows are created by the rotation of the earth acting on ocean water in confined basins.

The deep ocean circulation exists because of density differences and the tendency for stratification. The global circulation meanders through ocean basins as a result of topography. It is affected by the surface exchanges of heat and salinity. It is also affected by surface-driven forcing against boundaries that cause upwelling. We consider these complex interactions and their representation when we discuss the surface ocean (Sect. 6.3.4).

The ocean circulation is forced at large climate scales, but there is also variability of density, temperature, and topography at small scales. These small-scale differences lead to responses in the circulation. Some of these responses are very large

[3]Chlorofluorocarbons, or CFCs. The same chemicals that deplete the stratospheric ozone layer are inert in the absence of sunlight (in the ocean).

scale and depend mostly on the fundamental equations of motion and energy transfer, making them easier to represent in ocean models, but others are small scale and give rise to small-scale motions (**eddies**) throughout the ocean. These eddies may have important consequences for mixing ocean water properties that can ultimately affect the density and the circulation.

6.3.3 Eddies in the Ocean

Many of the critical processes that force the ocean, and the most uncertain ones, depend on variations at the small scale. For simulating the ocean, one of the most important problems is that internal wave motions and eddies travel slower because of smaller density gradients. This means that they maintain their coherence on smaller scales. The small size also means that current variations (ocean "weather systems") have a space scale that is smaller than the same critical scale in the atmosphere, and a longer timescale. The major oceanic flows, driven by the large-scale forcing of the wind, boundaries, and density variations of upwelling and downwelling, carry most of their energy in meandering small-scale eddies. This is a bit like the individual storm systems in the atmosphere that move large amounts of water at mid-latitudes.

Figure 6.4 shows ocean surface temperature variations in the **Gulf Stream** off the east coast of North America.[4] The Gulf Stream provides a vivid example of the complexity of ocean currents and eddies. The warm flow poleward is not a straight "highway" but meanders with swirls relating to instability on the sides of the current and interactions with the bottom topography. An idealized current might be like a drainage ditch with concrete sides (e.g., uniform lines in Fig. 6.2), whereas the actual ocean currents have lots of small-scale eddies like an uneven meandering stream (see Fig. 6.4). Eddies carry more of the energy in the ocean than they do in the atmosphere. As a result, ocean models are often run on finer grids than atmosphere models. A typical atmosphere model is 100–200-km resolution, and high resolution is 25 km, whereas a high-resolution ocean model would be 10 km, and standard resolution 25–100 km. Even so, these eddies are still often much finer scale than the grids in an ocean model, and representing how they form and evolve, and their effect on the flow, is a central problem of ocean modeling.

Ocean models contain several different types of eddies. As computation enables finer resolution, more of the mesoscale at 6–62 miles (10–100 km) can be resolved. But this means that representations of the sub-mesoscale from 0.6–6 miles (1–10 km) become more important. Because the motions are slower, and mix less than the atmosphere, the scales of motion in space become finer. More of the energy in the flow is contained in these structures. They are present throughout the depth of

[4]The image is of the surface temperature of the ocean from the Atmospheric Infrared Sounder (AIRS) satellite instrument. Public domain image credit: NASA.

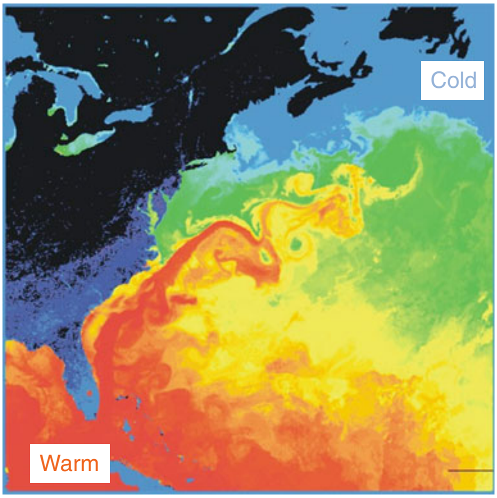

Fig. 6.4 Gulf stream. Satellite surface skin temperature of the North Atlantic from the atmospheric infrared sounder (AIRS). Coldest water in *blue*, warmest water in *red*; *orange*, *yellow*, and *green* are in between. The "Gulf Stream" of warm water from the Gulf of Mexico into the N. Atlantic is clearly seen with all of the eddies around it. *Image credit* National Aeronautics and Space Administration.

the ocean, but we usually see them only in the surface layer (such as in Fig. 6.4). Yet the large-scale effects of eddies mixing water are crucial for the largest climate scales (such as the flow in the Gulf Stream in Fig. 6.4). It is as if the large scales (think of the highway or the concrete drainage ditch) require the small scales to handle the flow (or energy) of the circulation. Currently, a great deal of ocean model development is focused on consistent representations of these eddies and their effects on large-scale mixing and overall circulation. It also drives ocean models to finer resolutions (6 miles or 10 km or even finer). Similar scale problems exist in the atmosphere with scales of motion that are within an order of magnitude or so of the grid scale: too coarse to resolve properly, but too fine to represent statistically.

6.3.4 Surface Ocean

The **surface ocean** is the primary region where the ocean communicates with the climate system. It is forced primarily by exchanges of energy, water, and momentum across the top boundary of the ocean. Whereas the deep ocean is driven by vertical gradients and vertical mixing, the surface ocean is mostly driven and affected by wind-driven forcing combined with topography (boundaries) and the effects of rotation. The forcing is highly variable and gives rise to eddies that comprise the large-scale flows. As evident in Fig. 6.4 and discussed earlier, representing the effects of these eddies in the ocean is a central problem of ocean modeling. These eddies are present in the wind-driven **gyres** (circular currents) we describe later in this section, and instabilities in the surface forcing can generate eddies. Furthermore, when water at the surface is pushed away from boundaries, it can cause upwelling of water from the deep ocean.

The surface ocean is forced by variations in atmospheric wind patterns. As with water or air that vertically wants to achieve its "correct" place in the column of density (stratification), atmospheric winds blow from high pressure to low pressure. At the surface, these pressures are coupled to patterns of energy input (which creates warmer or colder temperatures) and evaporation of water from the surface (which also can cool temperatures). Pushing water around an ocean basin causes currents, and it also causes slight differences in the height of the ocean as water "piles up" when forced against a boundary. The changes are too small to see, just a foot or less (10–25 cm) over thousands of miles, but stacking-up water means water wants to flow down this gradient. These forces act slowly and are affected by the earth's rotation. The good news is that the description of these forces can be well represented in equations in large-scale models. It is not a problem that the effect is small: It is representable because the effect occurs on a large scale. The ocean, of course, plays a large role in surface temperature and evaporation; hence, the atmosphere and ocean circulations are tightly coupled at the surface. This is true in many regions of the planet, from the tropics to the Arctic and the Antarctic.

From the perspective of the ocean, the wind induces a force on the ocean called a stress. **Wind stress** is visible in surface waves: The stronger the wind, the bigger the waves. This pushes the surface water, and the force is communicated through the body of the water column for some depth (decreasing as one gets deeper). The direction also changes with depth due to friction. The wind stress, combined with the boundaries in the ocean and the rotation of the earth, results in what is called **Sverdrup balance** (after a Norwegian oceanographer) between the force of the wind and the north-south transport of water. Water piles up on the west side of ocean basins in the tropics, and flows poleward, with an equatorward flow on the eastern side.

There is another feature of motion, however, and that is induced by rotation of the earth. Because the earth is rotating, there is an apparent sideways force, the **Coriolis force**, pushing to the right of motion in the Northern Hemisphere, and to the left of motion in the Southern Hemisphere.[5] As water flows to the west at the equator, this creates poleward motion. Thus, water comes from below along the equator to replace it (see Fig. 6.5). For water flow along a coast, the same force occurs. Equatorward return flow off the western part of continents (western North America and South America, and the Atlantic coast of Africa and Europe) induces

[5]Why is it in opposite directions when the earth spins the same way? The reason has to do with the angle of motion relative to the axis of the earth's rotation. For a complete description, see https://en.wikipedia.org/wiki/Coriolis_effect.

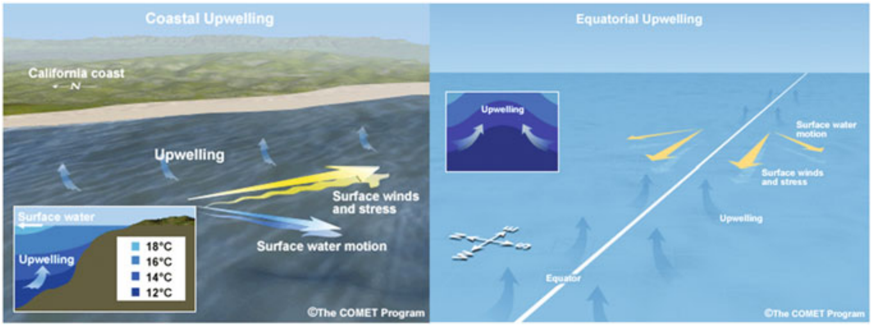

Fig. 6.5 Surface ocean. Coastal upwelling (*left panel*) due to southward surface winds along the coast, bringing up cold water. Equatorial upwelling (*right panel*), surface winds along the equator to the west (easterlies), causing water to move away from the equator and resulting in upwelling along the equator. Figure from the COMET program

offshore flow, also causing upwelling water from below to replace the surface water moving offshore. The Gulf Stream (see Fig. 6.4) is a manifestation of the poleward flow (the Kyushu current off of Japan is the Pacific version). So the flow induced by the large-scale rotation of the earth and the wind stress then has additional components from the basic physics of fluid on a rotating sphere.

The forces describe something called **geostrophic balance**: Water that is moving is affected by the earth's rotation, causing additional water motion. The oceanic surface gyres are driven by wind and the Coriolis force: The tropical westward flow of water in each hemisphere results from wind forcing along the equator (the trade winds). This induces a poleward Coriolis force that occurs on the west side of ocean basins. The eastward flow of water at mid-latitudes (again driven by prevailing winds) then induces an equatorward flow on the east side of basins.

The friction of ocean water combined with the Coriolis force also means that the force on ocean water is not in the same direction as the wind forcing (stress). If you push ocean water with wind, it tends to move a bit to the right of the direction of force (dictated by the earth's rotation). Because it is frictional, the layer below moves a bit more to the right, so that the net water motion is almost at a right angle to the wind stress. Along coastlines and the equator, this induces upwelling (illustrated in Fig. 6.5) from below as water flows away from a coast, or away from the equator. The cold upwelling water creates some of the temperature patterns in the surface ocean.

The mixing of momentum and heat down from the surface into the mixed layer is an important process for representing the structure of the ocean. Small-scale density gradients are induced by variations in temperature and salinity that make up this boundary layer at the top of the ocean. The ocean mixed layer is analogous to the atmospheric boundary layer above. Representing these fine-scale boundary layer processes in the ocean is important for coupling with the atmosphere above, not just for ocean circulations.

Exchanges of energy and water mass between the atmosphere and the surface of the ocean are also important. The ocean absorbs solar radiation, and some downward longwave (infrared) radiation from the atmosphere, and emits longwave radiation back. These exchanges help determine the temperature of the surface ocean. The exchange of freshwater with the surface is also a critical process. Water comes into the ocean from precipitation, rivers, or melting ice, and leaves by evaporation from the surface and formation of sea ice. Input and output water is fresh, so the addition or subtraction of fresh water with the same mass of salt will change the salinity and the density. Evaporation also is a cooling process, adding heat to the evaporated water, and removing it from the ocean. All of these exchanges of heat and mass of both water and salt must be accounted for exactly in an ocean model, and transferred to and from the atmosphere as appropriate. This is essentially a giant budget exercise. Like financial budgets track dollars, the surface energy and mass budgets have to track energy (watts) or mass (kilograms). The accounting has to be absolute: Even small systematic errors in mass and energy will be significant over long timescales.

6.3.5 Structure of an Ocean Model

The structure of an ocean model is internally similar to that of an atmosphere model. The ocean is divided into different grid cells distributed throughout the ocean. The ocean grids are often irregular and do not include points only on land. This makes them more complex. But ocean models at the same resolution have fewer points than atmosphere models, as they cover only 70 % of the earth's surface.

Ocean models also have a basic time-step loop. Typically, surface forcing from the atmosphere is calculated. This enables an estimate of the change in forcing on the ocean surface, and the change in pressure that will affect the height of the sea surface. When the atmospheric pressure drops, the ocean will tend to rise underneath it, and that water comes from somewhere else: inducing currents that need to be estimated.

Next, the different forcing terms on the ocean model, arising from different forces and parameterizations, are estimated. These include important parameterizations of eddies and eddy mixing. Changes to the mass of water and salt are estimated. One major difference between the atmosphere and the ocean is that, in the atmosphere, the forcing terms (clouds, radiative transfer) all occur independently in each column of the atmosphere: in one dimension in the vertical. In the ocean, the eddies mix horizontally as well as vertically, in three dimensions.

Finally, all these forcing terms for heat, currents, and even salinity are applied with the equations of motion for fluid (water) on a rotating sphere to get the resulting motions of water and changes to density in each grid location in the ocean. The tracers for chemicals in the ocean are updated. Then the revised state is iterated forward in time and the process begins again. This is very similar to how an atmosphere model works as described in Chap. 5.

6.3.6 Ocean Versus Atmosphere Models

There are many similarities between ocean and atmosphere models. Ocean models use most of the same scientific principles for fluid motion on a rotating sphere that apply to the atmosphere. There are difficulties in representing important mixing and transformation processes at small scales. Minor constituents (salt in the ocean, water in the atmosphere) play a major role in the general circulation. The grids and computational techniques of finite element modeling, and the use of subgrid-scale parameterizations in ocean models, are similar to those found in atmosphere models. But there are also major differences between the ocean and the atmosphere. The ocean has boundary layers on all surfaces (not just at the interface with the atmosphere). Ocean models have complex topography on both the bottom and the sides, and the ocean is effectively divided into basins (the five ocean basins of the Atlantic, Pacific, Arctic, Indian, and Southern Oceans). The ocean has more energy in the eddies and at smaller scales than the atmosphere, making their representation critical. Ocean models are forced by the atmosphere above, and that forcing is transmitted throughout the depth of the ocean, giving rise to eddies, surface currents, and the deep ocean circulation.

Thus far, we have focused on the physical description of the ocean. We return to some of the biogeochemical cycles in the ocean in Sect. 6.5.

6.4 Sea-Ice Modeling

Sea ice[6] is coupled closely with the ocean circulation. It concerns a representation of the freezing and melting process of ice and snow on top of the sea ice. Sea-ice models now also treat the dynamic motion of ice. Salt in (or expelled from) the ice is important for the ocean, and the ice strongly affects the flow of energy between the atmosphere and ocean. Sea ice forms in unique conditions at high latitudes where the temperature is cold. Ocean water freezes at about 28 °F (−2 °C) because of its salt content. Since temperatures vary strongly over the year, there is a large annual cycle in the extent of sea ice. In much of the polar regions, the sun has more of an annual cycle than a daily one, and above the Arctic and Antarctic circles, there are long periods when the sun is always present (“midnight sun”) or always absent (“polar night”).

[6]For a review of sea ice in the climate system, see Marshall, S. J. (2011). The *Cryosphere*. Princeton, NJ: Princeton University Press, Chap. 5.

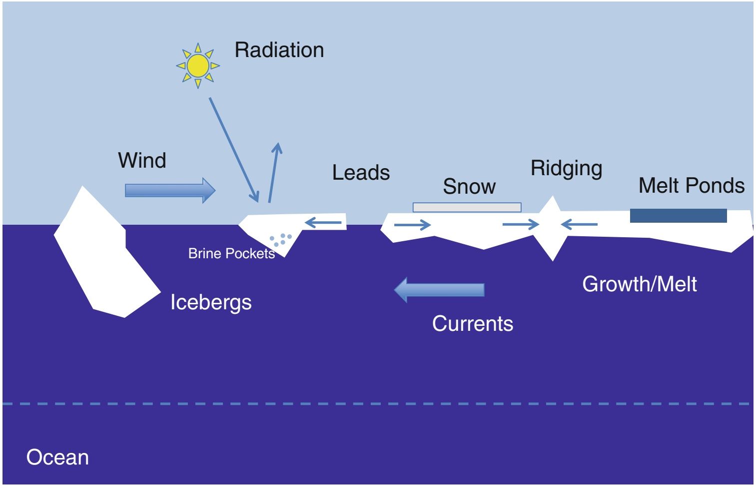

Fig. 6.6 Sea ice. Key processes in a sea ice model at the surface include winds, radiation, open areas between ice (leads), snow on ice and melt ponds. Key process at the base include currents, growth, melt and brine pockets

Sea-ice models must represent ice growth and melt. A schematic of many of the processes described is shown in Fig. 6.6. Models must account for the salt in seawater that is expelled back into the ocean, because this salt is critical for altering the density of the ocean at high latitudes. Along with temperature, it drives the formation of deep water at high latitudes. The salt content is dependent on the conditions of formation. Calmer water means more regular ice growth, and more salt expelled. In rougher conditions, **brine pockets** can become trapped in ice. Models must account for this inherently small-scale process to properly represent the exchange of salt and the density of the water underneath the ice.

Sea ice damps heat and moisture exchanges between ocean and atmosphere. This occurs because of the albedo contrast between bright ice and dark ocean and because ice is good at insulating the ocean from the atmosphere, reducing the surface exchange of heat, which must conduct through the ice, rather than convect (with density-driven motion) in the atmosphere. Convection is a lot more rapid than conduction (at least between ice, air, and water). So ice models must represent the conduction of heat. The heat conduction is dependent on the ice structure and can be altered by things like brine pockets and by the presence of snow on the ice. There is also a coupling of growth rate and thickness: Thin ice grows (and melts) faster because there is less resistance to conduction of heat through the ice to the atmosphere. Note that sea ice typically grows from the bottom, and the bottom may be irregular. In addition, snow falls onto the ice from the atmosphere, adding a little to the thickness. These are complicated processes in the thermodynamics of the evolution of sea ice.

In addition to the changes of state associated with ice formation and melting, the energy budget of sea ice must be accounted for. Radiation from the atmosphere, solar radiation (only in part of the year), and longwave radiation (all year) hit the ice. Ice and snow reflect some shortwave (solar) radiation, and absorb and emit longwave radiation. Sea-ice models must account for the radiative transfer in the top layers of ice and snow on the ice. As noted, the presence of snow on top of ice is quite important, as it changes the surface properties for both the absorption and scattering of light, as well as the conduction of heat (snow can "insulate" ice). The sea ice is a bit like the middle layer of a sandwich between the atmosphere and ocean, but it is critical for regulating exchanges between the two, and these exchanges have critical importance for how heat flows through the climate system.

These are thermal considerations, related to the conduction of heat through ice, its absorption, and the flux of heat to and from the atmosphere and ocean. The other critical aspect of sea-ice simulation is the dynamics, or motion, of sea ice. Sea ice is in constant motion, pushed by winds and by currents. The motion causes stresses in the ice and can cause it to deform. Sea ice is nearly flat, but with different layers that receive body forces from the ocean (bottom) and atmosphere (top). Thus, models must represent the momentum balance of the ice, the distribution of ice thickness, and the physics of the flow of ice in response to stresses (**rheology**). The thickness varies on small scales and is usually treated as a "thickness distribution" in any large-grid cell. The motion of the ice is predicted in response to the environment, stresses, and the internal structure of ice. Sea-ice motion can result in **leads** (open spaces) where there is divergent motion, and **ridging** in regions of convergent motion (see Fig. 6.6). Leads expose open water and often promote ice growth (or melt). Ridging increases ice thickness and helps ice survival: Thicker ice lasts longer. These processes are starting to be represented in sea-ice models. Many of these processes act on subgrid scales, occurring only in part of a grid box.

Sea-ice models act at the interface between the atmosphere and the ocean, and are subject to strong forcing from both. Traditionally they are often strongly coupled to ocean models and usually share the same grid as an ocean model. These ocean grids usually do not have a convergent "pole" in polar ocean regions, so the grid box sizes are usually nearly equal area.

As with the ocean, there are simplified sea-ice models. **Simplified sea-ice models** can assume a fixed distribution of sea ice (usually by month). More common is the use of thermodynamic considerations to estimate thickness and the local energy budget and energy fluxes. Mixed-layer ocean models are usually run with thermodynamic sea-ice models, since there are no ocean currents to move sea ice. A full sea-ice model adds a dynamic motion component to the simulation.

6.5 The Ocean Carbon Cycle

A new frontier of ocean modeling is the representation of the global cycle of carbon.[7] Carbon is a unique constituent of the climate system, since it passes through many of the different components: carbon dioxide (CO_2) as a gas in the atmosphere, into the land surface (as carbon-containing organic and inorganic matter), and also into and out of the ocean. CO_2 is dissolved in the water column, in chemical equilibrium with the atmospheric CO_2 pressure and ocean temperature. This enables ecosystems of aquatic plants to build biological matter (i.e., their bodies) with it, and forms the basis of the oceanic food chain. Carbon is present in calcium carbonate, which is also dissolved in water, and forms the shells of many marine organisms. Thus, when marine animals die, there is a steady buildup of carbon-rich sediments in the ocean. These oceanic carbon cycle processes are fundamental for affecting climate on long glacial and geologic timescales. They do not react very quickly (on timescales less than a century), but they may be important for understanding how ice ages occur, and how temperature and CO_2 vary with each other. The ocean is a vast store of carbon, and changing temperature and circulation may allow more or less carbon into the atmosphere, with resulting impacts on warming.

Models of the carbon cycle are beginning to be coupled with ocean and climate system models. They must represent different transformation processes for carbon based on organisms (biological carbon) and fundamental chemical processes (such as how much CO_2 is dissolved in seawater, which is a function of temperature). Some of these representations reflect simple chemical laws for how much CO_2 or carbonate is dissolved in seawater at a given temperature, and some are representations of biological processes. We treat the carbon cycle more fully in Chap. 7, on terrestrial systems.

6.6 Challenges

Ocean and sea-ice modeling is complex, and the different scales of motion (with small space scales, and very long timescales) pose a challenge for modeling. So, too, do the now-rapid changes in sea ice in the Northern Hemisphere. Some of these challenges are similar to the challenges faced by atmosphere models (such as the small scale of processes, and variability in a grid box), and some are unique to the ocean and ice.

[7] Archer, D. (2010). *The Global Carbon Cycle*. Princeton, NJ: Princeton University Press.

6.6.1 Challenges in Ocean Modeling

One major challenge in ocean modeling is a dearth of observations, particularly below the surface. The ocean is more difficult to observe than the atmosphere because it rapidly absorbs most wavelengths of radiation, and it is sparsely populated (by humans at least). So there are a lot less data beyond just the surface of the ocean. Most deep ocean data still have to be taken directly, which is not easy in a 4,000-m deep ocean. The rise of autonomous devices (buoys and small, unmanned undersea vehicles) enables remote measurements where unmanned submersible devices can rise and sink down to at least 2,000 m, recording temperature, salinity, and current measurements not unlike weather balloons in the atmosphere. Some have large fixed buoys at the top, and some drift, relaying their information to satellites when they surface. These systems are rapidly improving ocean observations and contributing to evaluation of ocean models, but observations of the deep ocean (below 2,000 m depth) are still very limited.

Another challenge in ocean modeling is properly representing the effect of small-scale eddies that cannot be explicitly simulated by a large-scale ocean model. Eddies move a lot of mass in the flow in the ocean, more like a meandering stream than a straight channel. This can be seen in the picture of small eddies in the Atlantic Gulf Stream in Fig. 6.4. Because the scale of eddies (6–30 miles, 10–50 km) starts approaching the ocean-model grid scale, it is difficult to represent them properly. Trying to represent a curvy flow in a stream is difficult if there are only one or two values for the current. One solution is "high-resolution" (6-mile or 10-km spacing) ocean models that "permit" the formation of eddies but are too coarse to resolve them properly.

Furthermore, ocean models have long adjustment timescales because of the deep ocean circulation. The use of simplified models of the mixed layer has come about since a full dynamic ocean model with a thermocline and a deep ocean circulation will reach a steady state (no change to climate with no external forcing) in about the time it takes for the ocean water to recycle, which is thousands of years of simulation. However, a mixed-layer model can reach equilibrium in only decades. The implication of this long timescale means that perturbations to the earth system will take thousands of years to equilibrate because of the slow processes in the ocean.

6.6.2 Challenges in Sea Ice Modeling

Sea-ice models have had quite a bit of recent success in simulating the observed distribution of sea ice (see Chap. 11). The comparison is complicated by the lack of observations of ice thickness. Since the arrival of full global weather satellite coverage in the 1970s, it is relatively easy to observe sea-ice coverage, but thickness observations remain elusive. The realism of ice thickness represented by sea-ice models is then uncertain. An additional complication is that the sea-ice distribution in sensitive seasons (like summer and fall) has been declining rapidly in the Arctic over the observational record, and it looks like the decline is getting faster (in terms of area of Arctic sea ice at a particular time, usually September). Most sea-ice models forced with atmospheric observations are able to reproduce the decline currently being experienced in the Arctic, but not necessarily the magnitude of the decline (they indicate less decline and more stable ice). Fully coupled models with an interactive ocean and atmosphere have a hard time reproducing the rapid decline of Arctic sea ice, likely due to the uncertainty in the forcing on the system going into these models and the complex interactions among ocean, ice, and atmosphere. It may be that the rapid sea-ice loss in the Arctic is a consequence of a long-term greenhouse warming signal, with short time period (i.e., a season or a year) additional variability. Because the sea ice is prone to melt and lasts from year to year, a set of events promoting loss in one year may result in lower ice the next year, and drive positive feedbacks. These surface feedbacks are discussed in Chap. 7.

6.7 Applications: Sea-Level Rise, Norfolk, Virginia

This case study demonstrates the roles of many stakeholders and the complex relationship between climate change science and other sources of information. Metropolitan Norfolk, Virginia, is a low-lying collection of cities and natural regions on the eastern coast of the United States. In addition to residential, recreational, and commercial activities, there is a large military presence, especially the U.S. Navy. Many private companies are associated with the naval presence, including unique dry-dock maintenance facilities.

Since 1971, flooding in Norfolk has increased from about 20 h a year to 130 h per year.[8] Between 1930 and 1997, only six storms brought **storm surges** (rising seas like a high tide due to wind and low atmospheric pressure) greater than 3 ft (1 m). Since 1997, there have been seven storms with surges greater than 3 ft. Though this part of the U.S. coast is often associated with hurricanes and tropical storms, wintertime Nor'easters are of equal importance when considering storm surges. This rapid increase in coastal flooding has sensitized the region to changes in sea level.

Analysis of the sea-level rise reveals that only part of it is due to the warming of the ocean and the melting of ice sheets and glaciers. The local land is sinking, partly because of rapid pumping of groundwater for residential and industrial use. In the past 100 years, the sinking land has been a larger effect than sea-level rise due to climate change. In addition to sea-level rise associated with the global average, there are strong local effects. These local effects are largely related to the variations

[8]Ezer, T., & Atkinson, Larry P. (2014). "Accelerated Flooding Along the U.S. East Coast: On the Impact of Sea-Level Rise, Tides, Storms, the Gulf Stream, and the North Atlantic Oscillations." *Earth's Future*, *2*: 362–382. doi:10.1002/2014EF000252.

in sea surface height associated with the Gulf Stream's being close to the coast (also see discussion in Chap. 8). Variations in the strength of the Gulf Stream cause variations in the "tilt" of the ocean surface, and thus the height at the coast. A strong Gulf Stream tilts the surface away from the coast, and some estimates predict a slowing of the Gulf Stream. The slowing would result in enhanced sea level rise along the coast. There are systematic variations of sea level in Norfolk associated with internal modes of variability in the atmosphere, for example, the North Atlantic Oscillation. Sea-level rise sits in context with internal variability and other causes of relative sea-level change, which is typical of many applications.

Partly because the increase in flooding is widely obvious and interferes with commerce and day-to-day life, sea-level rise has received much attention in what is generally a politically conservative region. Residents, businesses, cities, and the military are all active in developing sea-level-rise policy and plans. Local universities have performed research quantifying the different causes of sea-level rise and contributing to communication of what has happened, framing vulnerability, and what is likely to happen. Many nongovernmental organizations (NGOs) with different focuses represent particular interests ranging from conservation to social justice.

In the mid-2000s, the need to address sea-level rise and climate change became so apparent that a regional focus started to emerge, with the local Intergovernmental Pilot Project being formalized in 2014.[9] This organization strives to coordinate the sea-level-rise preparedness and resilience planning of federal, state, and local government agencies and the private sector and take into account the perspectives of the region's citizens.

Though sea-level rise projections are often viewed as highly uncertain, the convergence of observations, people's perception of vulnerability, effective communication, and concern for the viability of the region stand as motivation to take action. With planning periods focused on the next few decades, building standards, zoning, and codes are being modified. There is recognition that at the end of this planning horizon, sea-level rise will not be stable, but likely to be increasing. Therefore, sustained, future-looking planning and design will be required. Climate model results provide a range of estimates of possible future states to inform this process (though they may be uncertain; see Chap. 11). However, climate change effects on sea-level rise are only one part of the planning process (see Chap. 12).

6.8 Summary

The ocean and ice portions of the climate system have vastly different scales: Sea ice is a tenuous and thin layer that exists between the atmosphere and ocean, but plays a huge threshold role in regulating climate at high latitudes, and through

[9]Center for Sea Level Rise, http://www.centerforsealevelrise.org/.

albedo feedbacks, affecting the planet globally. The ocean itself is a huge reservoir of heat that damps changes to the forcing applied to it. The ocean can be divided into a surface ocean above the thermocline (containing the mixed layer) and the deep ocean. In the deep ocean, formation of deep water is critical. For the mixed layer, surface fluxes and mixing are critical. Mixing occurs on smaller scales in the ocean than in the atmosphere, so ocean models are often run at finer scales, and mixing processes need to be well represented. Finally, the ocean supports a significant ecosystem (or set of ecosystems) that affect the carbon dioxide dissolved in the ocean and transferred either back to the atmosphere or into deep ocean sediments. The ocean is the link between the "fast" climate system (decades to centuries) and "geologic" timescales (millennia to millions of years).

Key Points

- The ocean is stratified, and density is important. Heat and salt control density.
- Ocean currents are driven by surface winds and the rotation of the earth, and deep currents are driven by density.
- Ocean models must represent small-scale eddies.
- Sea-ice models can represent recent losses in Arctic ice.
- The ocean carbon cycle is important for the global carbon cycle.