Chapter 8

Storm skies over the ocean.

Circulation and Ocean Structure

Chapter Outline

Learning Outcomes

After studying the information in this chapter students should be able to:

1. *estimate* the density of a mixture of two samples of seawater that have the same density but different temperatures and salinities,
2. *describe* and *sketch* changes in the seasonal thermocline at mid-latitudes through the year,
3. *plot* temperature and salinity as a function of depth and *identify* the thermocline and halocline,
4. *list* five different water masses and *describe* how they form,
5. *relate* surface convergence and divergence to downwelling and upwelling,
6. *describe* the properties of water masses in each ocean basin, and
7. *discuss* how thermal energy can be extracted from the oceans.

Hidden below the ocean's surface is its structure. If we could remove a slice of ocean water in the same way we might cut a slice of cake, we would find that, like a cake, the ocean is a layered system. The layers are invisible to us, but they can be detected by measuring the changing salt content and temperature and by calculating the density of the water from the surface to the ocean floor. This layered structure is a dynamic response to processes that occur at the surface: the gain and loss of heat, the evaporation and addition of water, the freezing and thawing of ice, and the movement of water in response to wind. These surface processes produce a series of horizontally moving layers of water, as well as local areas of vertical motion. In this chapter, we will study both the surface processes and their below-the-surface results in order to understand why the ocean is structured in this way and how the structure is maintained. We will also explore the ways in which oceanographers gather data about this layered system, and we will survey the possibilities of extracting useful energy from it.

8.1 Density Structure

Surface Processes

Absorption of solar energy and energy exchanges between the sea surface and the atmosphere determine sea-surface temperatures and the zones of precipitation and evaporation that govern surface salinities. Variations in temperature and salinity combine to control the density of ocean surface water. Many combinations of salinities and temperatures produce the same density; the result is shown in lines of constant density (fig. 8.1). Figure 8.1, called a T-S diagram, is related to the density values in figure 5.8 in chapter 5.

As the salinity increases, the density increases; as the temperature increases, the density decreases. Salinity may be increased by evaporation or by the formation of sea ice; it may be decreased by precipitation, the inflow of river water, the melting of ice, or a combination of these factors. Changes in pressure also affect density. As the pressure increases, the density increases. However, because pressure plays a minor role in determining the density of surface water, its effects are not considered here.

Notice that in figure 8.1 the lines of constant density are curved. Therefore, two types of water having the same density but different values of salinity and temperature when mixed form a water that lies on their mixing line. This water has a density greater than either of the original water types, and it will sink. This mixing and sinking process is known as **caballing** and occurs in all oceans when surface waters converge.

Less-dense water remains at the surface (for example, the warm, low-salinity surface water of the equatorial latitudes). Although the surface water at 30°N and 30°S latitudes is warm,

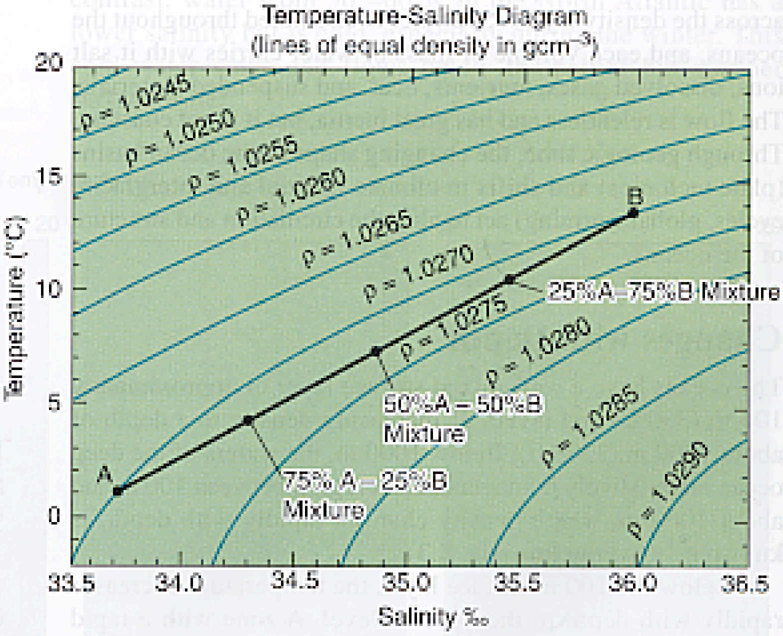


Figure 8.1 The density of seawater, measured in grams per cubic centimeter, is abbreviated as ρ (rho) and varies with temperature and salinity. Many combinations of salinity and temperature produce the same density. Low densities are at the *upper left* and high densities at the *lower right*. The *straight line* is the mixing line for waters *A* and *B*, both with the same density. A mixture of *A* and *B* lies on the mixing line and is more dense than either *A* or *B*.

it has a higher salinity than surface waters at the equator. Therefore, it is denser than the warm, low-salinity equatorial water. This 30° latitude surface water sinks below the equatorial surface water; it extends from the surface at 30°N to below the less-dense equatorial layer and back to the surface at 30°S. The combined salinity and temperature in surface waters at 50°–60°N and 50°–60°S produce a water that is denser than either the equatorial or the 30° latitude surface water. The 50°–60° latitude water therefore sinks below the equatorial and 30° surface water and extends from the surface in one hemisphere below the other water types to the surface in the other hemisphere. Winter conditions in the subpolar regions lower the surface water temperature and, if sea ice forms, increase the salinity. The result is a dense surface water that sinks at the subpolar latitudes. These variations in surface water properties and the resulting density changes produce a density-layered ocean. This layered system is shown in figure 8.2. The thickness and horizontal extent of each layer are related to the rate at which the water of each layer is formed and the size of the surface region over which it is formed.

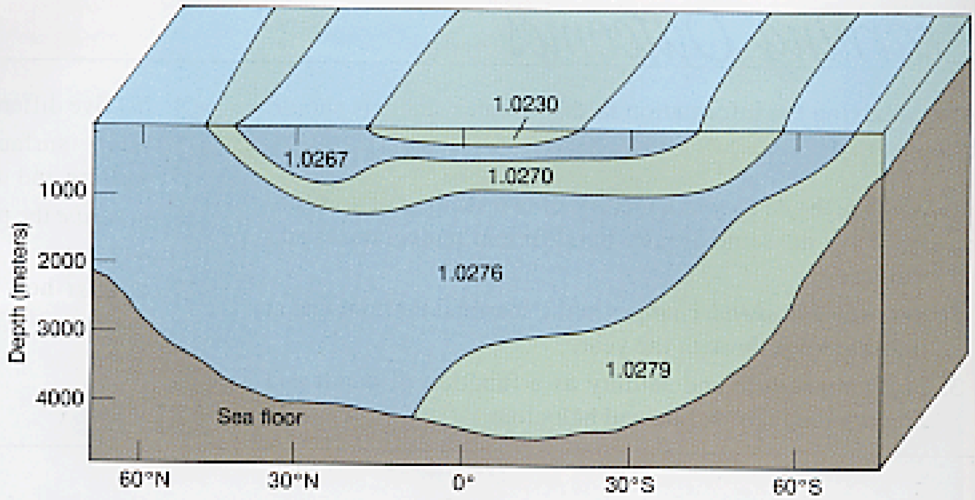


Figure 8.2 Waters of different densities form a layered ocean. The density of each layer is determined at the surface by the climate at the latitude at which it is formed. Density values are given in grams per cubic centimeter.

Wind-driven surface currents also directly affect the density structure of the upper ocean and continually move surface waters from one area to another. Turbulence associated with tidal motion as well as wind-driven surface currents mix water across the density layers. Seawater is circulated throughout the oceans, and each volume or mass of water carries with it salt ions, dissolved gases, nutrients, heat, and suspended materials. The flow is relentless and has great inertia, but it is not constant. Through geologic time, the changing shape of the ocean basins (plate tectonics) and shifts in climate (glacial and interglacial cycles, global warming) act to alter the circulation and structure of the oceans.

Changes with Depth

The oceans have a well-mixed surface layer of approximately 100 m (330 ft) and layers of increasing density to a depth of about 1000 m (3300 ft). Below 1000 m, the waters of the deep ocean are relatively homogeneous. A region between 100 m and about 1000 m, where density changes rapidly with depth, is known as a **pycnocline** (fig. 8.3).

Below the 100 m surface layer, the temperature decreases rapidly with depth to the 1000 m level. A zone with a rapid change in temperature with depth is called a **thermocline.** Below the thermocline, the temperature is relatively uniform over depth, showing a small decrease to the ocean bottom. A similar situation occurs with salinity. Below the surface water at the middle latitudes, the salinity increases rapidly to about 1000 m; this zone of relatively large change in salinity with depth is called the **halocline.** Beneath the halocline, relatively uniform conditions extend to the ocean bottom. Both a thermocline and a halocline are shown in figure 8.4.

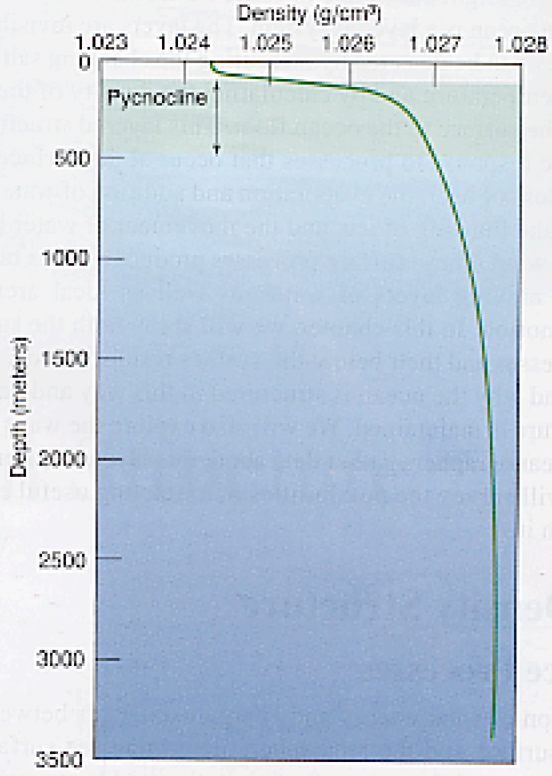


Figure 8.3 Density increases with depth in seawater. The pycnocline is the region in which density changes rapidly with depth. (Based on data from the northeastern Pacific Ocean.)

If the density of the water increases with depth, the water column from surface to depth is **stable.** If there is more-dense water on top of less-dense water, the water column is **unstable.** An unstable water column cannot persist; the denser surface water sinks and the less-dense water at depth rises to replace the surface water. Vertical convective **overturn** of the water takes place. If the water column has the same density over depth, it has neutral stability and is termed **isopycnal.** A neutrally stable water column is easily vertically mixed by wind, wave action, and currents. If the water temperature is unchanging over depth, the water column is **isothermal;** if the salinity is constant over depth, it is **isohaline.**

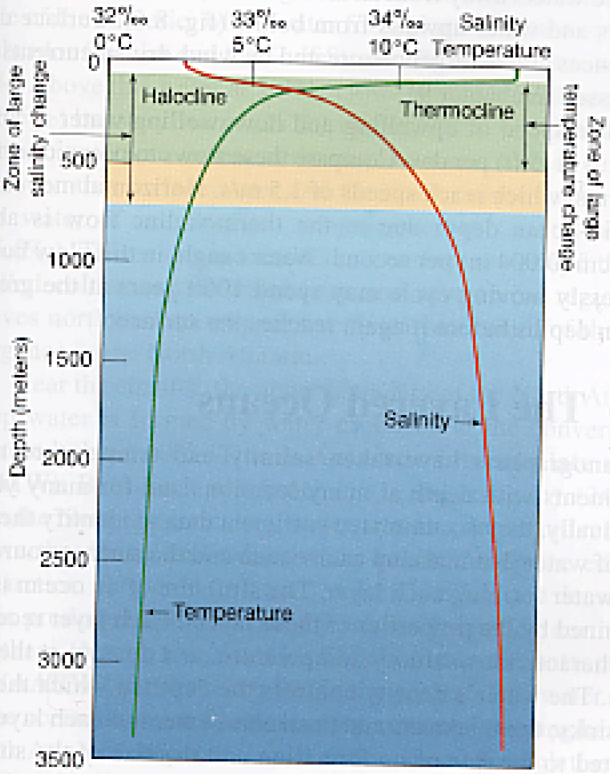


Figure 8.4 Temperature and salinity values change with depth in seawater. Rapid changes in temperature and salinity with depth produce a thermocline and a halocline, respectively. (Based on data from the northeastern Pacific Ocean.)

Density-Driven Circulation

Processes that increase the water's density at the surface cause density-driven **vertical circulation.** This overturn may reach only to shallow depths or extend to the deep-sea floor, ensuring an eventual top-to-bottom exchange of water. Because the density is normally controlled by surface changes in temperature and salinity, this vertical circulation is often called **thermohaline circulation.** An excellent example of thermohaline circulation occurs in the Weddell Sea of Antarctica, where winter cooling and freezing produce dense, cold, saltier surface water that sinks to the sea floor. This water descending along the coast of Antarctica is the densest water found in the open oceans (see fig. 8.2).

At the temperate latitudes in the open ocean, the surface water's temperature changes with the seasons. This effect is illustrated in figure 8.5. During the summer, the surface water warms and the water column is stable, but in the fall, the surface water cools, its density increases, and overturn begins. Winter storms and winter cooling continue the mixing process. The shallow thermocline formed during the previous summer is lost, and the upper portion of the water column is vertically mixed and becomes isothermal to greater depths. Spring brings warming, and the shallow thermocline begins to reestablish itself; the water column becomes stable and remains so through the summer.

Seasonal temperature changes are more important than salinity changes in altering density in the open ocean at subarctic and high temperate latitudes. For example, in the Atlantic Ocean, the surface water at 30°N and 30°S has a high salinity but is warm year-round, so it stays at the surface. In contrast, water from 50°–60°N in the North Atlantic has a lower salinity but is cold, especially during the winter. This cold water sinks and flows below the saltier but warmer surface water.

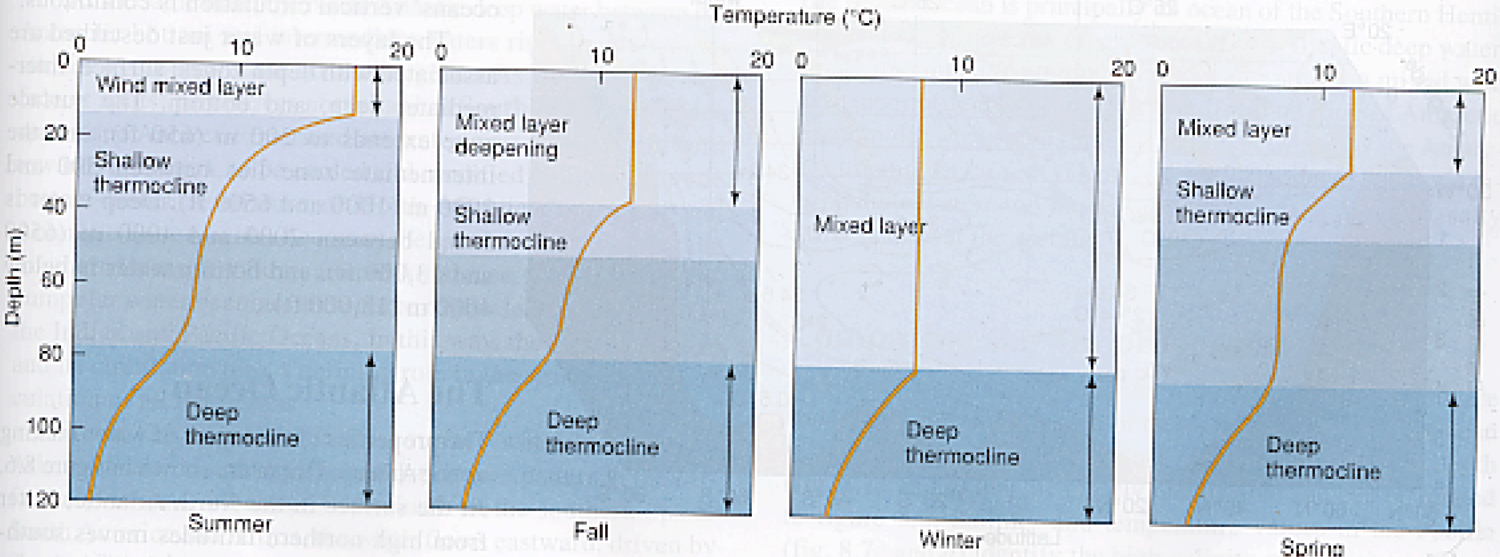


Figure 8.5 The surface-layer temperature structure varies over the year. In the absence of strong winds and wave action in the summer, solar heating produces a shallow thermocline. During the fall and winter, the surface cooling and storm conditions cause mixing and vertical overturn, which eliminate the shallow thermocline and produce a deep wind-mixed layer. In spring, the thermocline re-forms. The deep thermocline is below the influence of seasonal surface changes and remains in place year-round.

Close to shore, the salinity of seawater can become more important than the temperature in controlling density. Salinity is particularly important in semienclosed bays, sounds, and fjords that receive large amounts of freshwater runoff. Here, extremely cold (0°–1°C) fresh water from melted ice is added to the sea. In some cases, the salinity becomes so low that the diluted seawater does not sink but remains at the surface as a seaward-moving **freshwater lid.** In polar regions, when sea ice melts, a layer of fresh water forms at the surface and slowly dilutes the underlying surface water.

8.2 Upwelling and Downwelling

When dense water from the surface sinks and reaches a level at which it is denser than the water above but less dense than the water below, it spreads horizontally as more water descends. At the surface, water moves horizontally into the region where sinking is occurring. The dense water that has descended displaces deeper water upward, completing the cycle. Because water is a fixed quantity in the oceans, it cannot be accumulated at one location or removed at another location without movement of water between those locations. This concept is called **continuity of flow.** Areas of thermohaline circulation where water converges and sinks are called **downwelling zones;** areas of diverging rising waters are **upwelling zones.** Downwelling is a mechanism that transports oxygen-rich surface water to depth, where it is needed for the deep-living animals. Upwelling returns low oxygen-content water with dissolved, decay-produced nutrients to the surface, where the nutrients act as fertilizers to promote photosynthesis and the production of more oxygen in the sunlit surface waters.

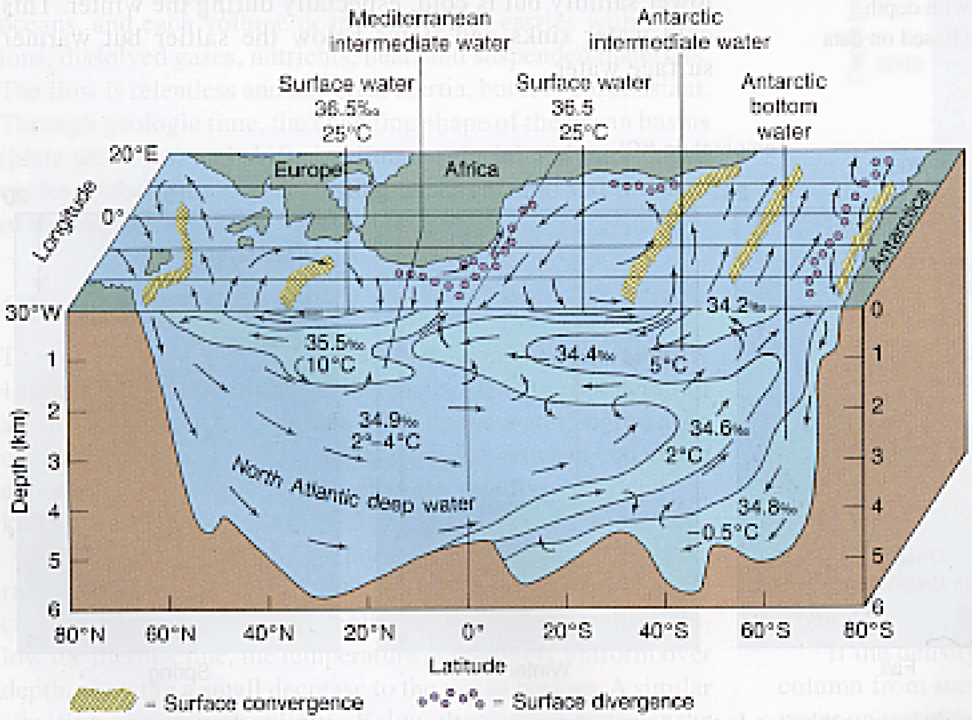


Figure 8.6 The anatomy of the Atlantic Ocean. Surface and subsurface circulation are related here. The surface currents converge to produce downwelling; water sinks to its density level and flows horizontally. Water at depth rises under zones of surface divergence.

Upwelling and downwelling refer to vertical motion of water upward or downward. They are present in thermohaline circulation but can also be caused by wind-driven surface currents. When the surface waters are driven together by the wind or against a coast, a surface **convergence** is formed. Water at a surface convergence sinks, or downwells. When the wind blows surface waters away from an area or coast, a surface **divergence** occurs and water upwells from below (fig. 8.6). Surface convergences and divergences created by wind-driven currents are discussed in chapter 9.

The speed of upwelling and downwelling water is about 0.1–1.5 m (5 ft) per day. Compare these flows to oceanic surface currents, which reach speeds of 1.5 m/s. Horizontal movement at mid-ocean depth due to the thermohaline flow is about 0.01 cm (0.004 in) per second. Water caught in this slow but relentlessly moving cycle may spend 1000 years at the greater ocean depths before it again reaches the surface.

8.3 The Layered Oceans

Oceanographers have taken salinity and temperature measurements with depth at many locations and for many years. Gradually, they accumulated sufficient data to identify the layers of water that make up each ocean and the surface source of the water forming each layer. The structure of an ocean is determined by the properties of these layers. Each layer received its characteristic salinity, temperature, and density at the surface. The water's density controls the depth to which the water sinks; the thickness and horizontal extent of each layer are related to the rate of its formation and the size of the surface source region. Water that sinks from the surface to spread out at depth and slowly mixes with adjacent layers eventually rises at another location. In all cases, water that sinks displaces an equivalent volume of water upward toward the surface at some other location so that the oceans' vertical circulation is continuous.

The layers of water just described are associated with depth zones: surface, intermediate, deep, and bottom. The surface zone extends to 200 m (650 ft), and the intermediate zone lies between 300 and 2000 m (1000 and 6500 ft). Deep water is found between 2000 and 4000 m (6500 and 13,000 ft), and bottom water is below 4000 m (13,000 ft).

The Atlantic Ocean

The properties of the layers of water making up the Atlantic Ocean are shown in figure 8.6. At the surface in the North Atlantic, water from high northern latitudes moves southward, while water from low latitudes moves northward along the coast of North America and then east across the North Atlantic. These waters converge in areas of cool temperatures and high precipitation at approximately 50°–60°N in the Norwegian Sea and at the boundaries of the Gulf Stream and the Greenland and Labrador Currents. The resulting mixed water has a salinity of about 34.9‰ and a temperature of 2°–4°C. This water, known as **North Atlantic deep water (NADW),** sinks and moves southward. North Atlantic deep water from the Norwegian Sea moves south along the east side of the Atlantic, while water formed at the boundary of the Labrador Current and the Gulf Stream flows along the western side. Above this water at 30°N, a low-density lens of very salty (36.5‰) but very warm (25°C) surface water remains trapped by the circular movement of the major oceanic surface currents. Between this surface water and the North Atlantic deep water lies water of intermediate temperature (10°C) and salinity (35.5‰). This water is a mixture of surface water and the upwelled colder, saltier water from the subtropical regions. It moves northward to reappear at the surface south of the convergence in the North Atlantic.

Near the equator, the upper boundary of the North Atlantic deep water is formed by water produced at the convergence centered about 40°S. This is **Antarctic intermediate water (AAIW).** Because it is warmer (5°C) and less salty (34.4‰) than the North Atlantic deep water, it is less dense and remains above the denser and saltier water below. Along the edge of Antarctica, very cold (−0.5°C), salty (34.8‰), and dense water is produced at the surface by sea ice formation during the Southern Hemisphere's winter. This is **Antarctic bottom water (AABW),** the densest water in the oceans. This water sinks to the ocean floor and flows slowly northward, creeping beneath North Atlantic deep water, as it continues on through the deep South Atlantic ocean basins west of the Mid-Atlantic Ridge. Antarctic bottom water does not accumulate enough thickness to be able to flow over the mid-ocean ridge system into the basins on the African side of the ridge; it is confined to the deep basins on the west side of the South Atlantic and has been found as far north as the equator.

At the same time, the North Atlantic deep water between the Antarctic bottom and intermediate waters rises to the ocean's surface in the area of the 60°S divergence. As it reaches the surface, it splits; part moves northward as **South Atlantic surface water** and Antarctic intermediate water; part moves southward toward Antarctica, to be cooled and modified to form Antarctic bottom water. A mixture of North Atlantic deep water and Antarctic bottom water becomes the circumpolar water for the Southern Ocean that flows around Antarctica. The Antarctic circumpolar water becomes the source of the deep water found in the Indian and Pacific Oceans. In this way, the Atlantic Ocean and its circulation play a defining role in the structure and circulation of all the oceans.

Warm (25°C), salty (36.5‰) surface water in the South Atlantic is also caught by the circular current pattern at the surface and is centered about 30°S. South of the southern tips of South America and Africa, the water flows eastward, driven by the prevailing westerly winds, which move the water around and around Antarctica.

Because the Atlantic Ocean is a narrow, confined ocean of relatively small volume but great north-south extent, the water types are readily identifiable and the movement of the layers can be followed quite easily. In addition, the bordering nations of the Atlantic have had a long-standing interest in oceanography, so the vertical circulation and layering of the Atlantic are the most studied and the best understood of all the oceans.

The Pacific Ocean

In the vast Pacific Ocean, waters that sink from relatively small areas of surface convergences lose their identity rapidly, making the layers difficult to distinguish. Antarctic bottom water forms in small amounts along the Pacific rim of Antarctica, but it is quickly lost in the great volume of the Pacific Ocean. The deeper water of the South Pacific Ocean is the water of the Antarctic circumpolar flow. Because the North Pacific is isolated from the Arctic Ocean, only a small amount of water comparable to North Atlantic deep water can be formed. In the extreme western North Pacific, convergence of the southward-flowing cold water from the Bering Sea and the Sea of Okhotsk and the northward-moving water from the lower latitudes produces only a small volume of water that sinks to mid-depths. There is no large source of deep water similar to that found in the North Atlantic. Warm, salty surface water occurs at subtropical latitudes (30°N and 30°S) in each hemisphere, and Antarctic intermediate water is produced in small quantities, but its influence is small. Deep-water flows in the Pacific are sluggish, and conditions are very uniform below 2000 m (6600 ft). The slow circulation of the Pacific means that it has the oldest water at depths where age is measured as time from the water's last contact with the surface. Residence time for deep water in the Pacific is about twice that of deep water in the Atlantic.

The Indian Ocean

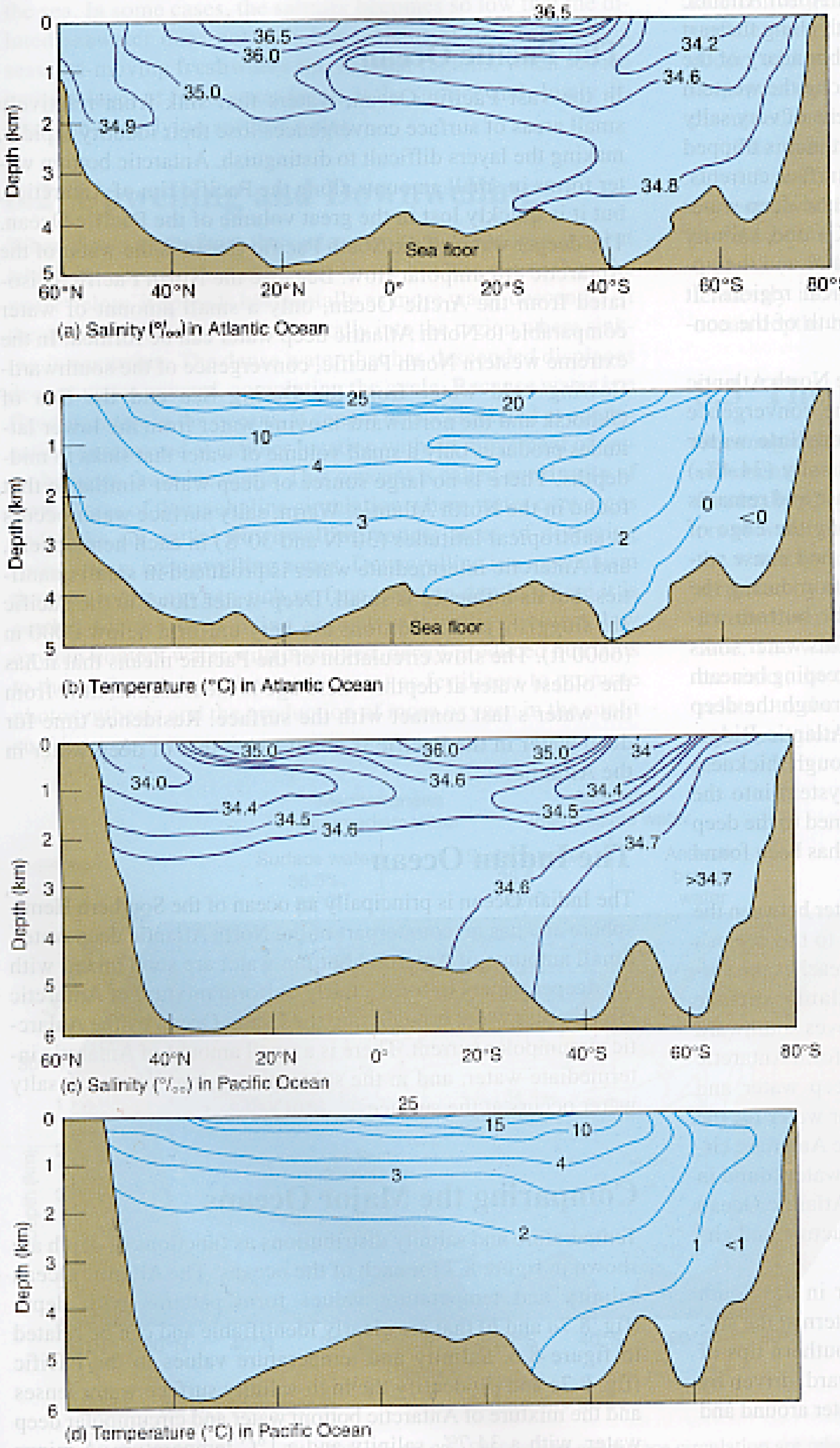
The Indian Ocean is principally an ocean of the Southern Hemisphere and has no counterpart of the North Atlantic deep water. Small amounts of Antarctic bottom water are soon mixed with the deeper waters to form a fairly uniform mixture of Antarctic circumpolar water brought into the Indian Ocean by the Antarctic circumpolar current. There is a small amount of Antarctic intermediate water, and in the subtropics, a lens of warm, salty water occurs at the surface.

Comparing the Major Oceans

Temperature and salinity distributions as functions of depth are shown in figure 8.7 for each of the oceans. The Atlantic Ocean salinity and temperature values form patterns with depth (fig. 8.7*a* and *b*) that are clearly identifiable and can be related to figure 8.6. Salinity and temperature values in the Pacific (fig. 8.7*c* and *d*) identify the high-salinity surface water lenses and the mixture of Antarctic bottom water and circumpolar deep water, with a 34.7‰ salinity and a 1°C temperature. A minor intrusion of low-salinity water in the north is the result of the surface convergence in the far northwestern Pacific. Note the large volume of deep water that shows a uniform salinity in the Pacific Ocean. Temperature values also follow a less-definite pattern than in the Atlantic, emphasizing the uniformity of most of this deep water. The Indian Ocean values (fig. 8.7*e* and *f*) resemble those for the South Atlantic, without the presence of water that is comparable to North Atlantic deep water. The Arctic Ocean is an extension of the North Atlantic; see figure 8.8 and the following discussion.

The Arctic Ocean

Some oceanographers consider the Arctic Ocean to be an extension of the North Atlantic, but it differs from the Atlantic Ocean in many significant ways. About one-third of its area, 8×10^6 km^2 (3.1×10^6 mi^2), is covered by extensive continental shelves, the widest of any ocean. Two basins, the Eurasian to the east and the Canadian to the west, occupy the central portion of the ocean; they are separated by the Lomonosov Ridge extending due north from Greenland (fig. 8.8). The Eurasian basin is the deeper basin, 5000 m (16,400 ft); it is connected to the North Atlantic through a gap in the continental shelf between Spitsbergen and Greenland. The larger, shallower Canadian basin is about 3800 m (12,450 ft) deep. Both basins contain spreading centers that are northern extensions of the Mid-Atlantic Ridge system.



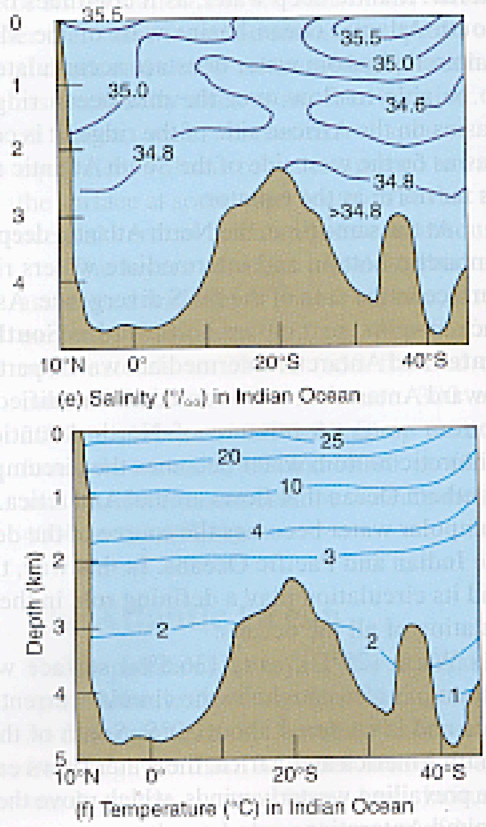


Figure 8.7 Mid-ocean salinity and temperature profiles of the Atlantic, Pacific, and Indian Oceans. Temperature and salinity are shown as functions of depth and latitude.

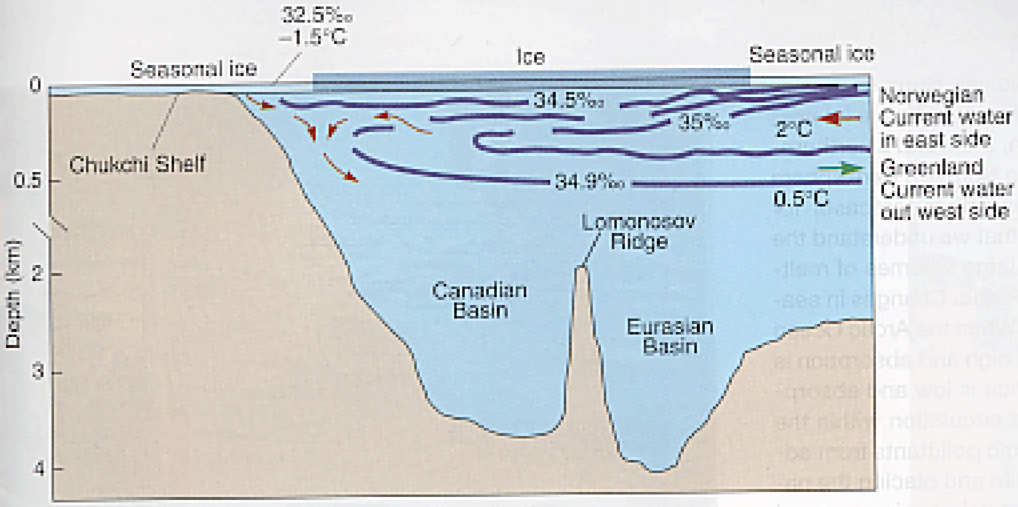


Figure 8.8 The Arctic Ocean. Water is supplied to a shallow surface layer by the Atlantic Ocean's Norwegian Current. Water exits the Arctic Ocean into the Atlantic Ocean by the Greenland Current. The circulation is confined almost entirely to the upper 500 m (1650 ft) of the Arctic Ocean.

The density of the Arctic Ocean water is controlled more by salinity than by temperature. Its surface layer is formed from low-salinity water entering from the Bering Sea, fresh water from Siberian and Canadian rivers, and seasonal melting of sea ice. The surface layer from these combined sources is about 80 m (250 ft) deep and has a low salinity (32.5‰) and a low temperature (−1.5°C). Below the surface layer, salinity increases with depth in the halocline layer, 200 m (650 ft) thick, to reach 34.5‰ at its base. The cold, salty water of the halocline layer is produced by the annual freezing and formation of sea ice over the continental shelves. This water sinks and moves across the shelves to spread out in the central ocean basins. West of Spitsbergen, North Atlantic water (2°C and 35‰) enters the Arctic Ocean and is cooled as it flows under the halocline and fills the Arctic Ocean basins. This water upwells along the edge of the continental shelves, mixing with the water formed during freezing, and exits the Arctic as water of 0.5°C temperature and 34.9‰ salinity along the edge of the shelf adjacent to Greenland. This exiting water moves south along the coast of Greenland and enters the North Atlantic south of Greenland and Iceland, where it combines with Gulf Stream water to form North Atlantic deep water.

Many aspects of the structure of the Arctic Ocean are still a puzzle. Approximately 70% of the Arctic Ocean observations to date were made by the former Soviet Union. Since the cessation of the Cold War, detailed bathymetric and water column data have been declassified and made available to the Arctic research community. One data set contained 1.3 million salinity and temperature measurements taken over a period of many years from icebreakers, drifting ice camps, and buoys. The data are being distributed in a four-volume *U.S.-Russian Arctic Atlas*. The first volume was released in January 1997.

The average minimum Arctic Sea ice cover since 1978 has been 6.22×10^6 km^2 (2.4×10^6 mi^2). In 2002, measurements of Arctic Ocean ice cover showed the summer minimum to have decreased to 5.18×10^6 km^2 (2×10^6 mi^2). The sea ice that survived the summer of 2002 was also found to be thinner than usual. Ice reflects solar radiation, but exposed water absorbs solar radiation and so accelerates the melting process. If the present situation continues, an estimated 20% of ice cover will be lost by 2050, which could lead to problems for polar bears, seals, and natives of the high northern latitudes. Under such conditions, Greenland's glaciers would also melt and release large amounts of fresh water to the North Atlantic. Such water would both cool and reduce the density of North Atlantic surface water. In these circumstances, the production of North Atlantic deep water and the circulation of the Atlantic would decrease.

Bordering Seas

Two small but specific water types from bordering seas are readily identifiable, one in the North Atlantic and one in the Indian Ocean. The water from the Mediterranean Sea has a temperature of about 13°C and a salinity of 37.3‰ as it leaves the Strait of Gibraltar. This water, mixing with Atlantic Ocean water, forms an intermediate density water, **Mediterranean Intermediate Water (MIW)**, that sinks in the North Atlantic to a depth of approximately 1000 m (3300 ft) (fig. 8.6). The influence of Mediterranean water can be traced 2500 km (1500 mi) from the Strait of Gibraltar before it is lost through modification and mixing. In the Indian Ocean, the initially very salty (40–41‰) water from the Red Sea has been found in a spreading layer at 3000 m (10,000 ft) depth more than 200 km (124 mi) south of its source.

Studies to determine where and how the deep water of the Mediterranean is formed were made in the 1980s and again in 1995. The Mediterranean Sea is divided into east and west basins at the Strait of Sicily (fig. 8.9). Deep and intermediate Mediterranean waters formed in the east basin move through the Strait of Sicily into the west basin and then through the Strait of Gibraltar into the Atlantic Ocean. The 1980 studies showed that high evaporation rates and winter cooling in the southern Adriatic Sea caused surface water to sink, and this water then moved into the east basin and from there into the west basin on its way to the Atlantic. The 1995 studies indicate that the source of the east basin water is no longer the Adriatic Sea; instead, the water is being formed in the Aegean Sea. The Aegean Sea water is saltier and denser than that produced in the Adriatic Sea, and the Aegean water is now displacing older Adriatic water upward and westward into the Atlantic Ocean. It appears that small climate changes in the eastern Mediterranean Sea affect winter cooling and evaporation rates and play a significant role in the location and strength of the processes that produce the deep waters of the Mediterranean. Recent studies of surface currents show that Atlantic Ocean water flows

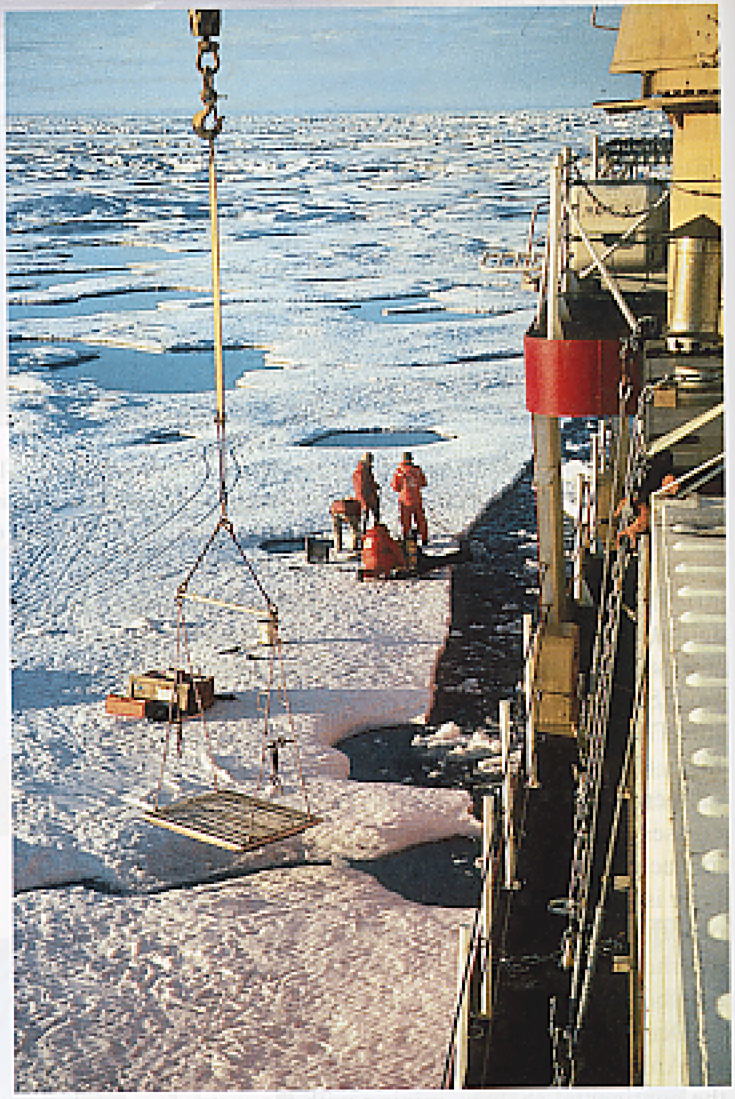
Arctic Ocean Studies

Not all oceanographic research involves ships or is conducted in warm, tropical regions. More and more interest is being shown in the Arctic Ocean, the world's northernmost but least-studied ocean. Understanding Northern Hemisphere climate changes and making long-range weather forecasts for northern Europe and North America require that we understand the effects of periodic increases in ice and the large volumes of meltwater that enter the North Atlantic from the Arctic. Changes in sea-ice cover directly affect Earth's heat budget. When the Arctic Ocean is ice-covered, reflectance of solar energy is high and absorption is low, but when little ice is exposed, reflectance is low and absorption is high. New concerns have arisen that circulation within the Arctic Ocean may distribute persistent organic pollutants from adjacent land sources, contaminating marine life and placing the native peoples dependent on this marine life at risk; again, we need more information on the circulation of the Arctic Ocean to evaluate this problem.

In 1994, a joint program between Canada and the United States used icebreakers to traverse 1600 nautical miles of the Arctic Ocean from the Bering Sea to the pole. The main goal was to establish the Arctic Ocean's role in global climate change. Studies were made of the water, sea ice, sea floor, marine organisms, distribution of pollutants, and polar bear populations (box figs. 1 and 2). The overlying theme was to establish the Arctic Ocean's role in global climate change by studying the area from the atmosphere through the water and ice down to the sea floor.

On January 1, 1994, the Arctic Climate System Study (ACSYS) was organized. The observational phase of this program continued for ten years. Its primary goal was to understand the role of the Arctic in producing global climate. ACSYS encouraged and coordinated national and international research activities concerning ocean circulation, ice cover, exchanges of water in the Arctic Ocean, long-term climatic research, and monitoring programs. The Surface Heat Budget of the Arctic (SHEBA) was a subprogram of ACSYS, as well as a U.S.-Canadian cooperative research program. This program froze a Canadian Coast Guard icebreaker into the Arctic ice in October 1997, 248 km (400 mi) north of Prudhoe Bay, Alaska. For the next year, the ship acted as hotel and supply base for the SHEBA program (box fig. 3). The data from studies of air-sea exchange processes affecting Arctic Ocean heat budgets helped scientists better understand the role of Arctic processes in global climate models.

In the spring of 2000, researchers began a new program known as the National Science Foundation's North Pole Environmental Observatory. To learn more about how the northernmost ocean regulates global climate, buoys were placed on the polar ice to drift with the ice pack. These buoys are equipped with sensors that extend through the ice and gather data on changes in the thickness of the ice and the properties of the upper ocean. Information is relayed via satellites. Another installation of drifting buoys was completed in the spring of 2001. One buoy was developed by the Japanese to measure ocean temperature and salinity, current profiles, and atmospheric temperature and pressure, as well as wind velocity. Other buoys include a meteorological buoy that records wind speed, temperature, and



Box Figure 1 Scientists and their gear are offloaded to the ice from the U.S. Coast Guard icebreaker *Polar Sea* during its 1994 Arctic Ocean Section expedition.

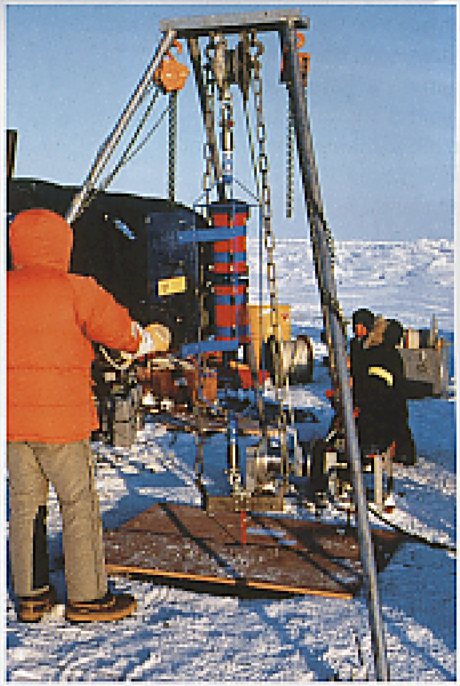
atmospheric pressure; two radiometer buoys that measure radiation from the sun and reflected radiation from the ice; and two ice-mass balance buoys that record ice temperature and snow thickness.

Another installation in 2001 was a 4250 m (14,000 ft) long instrumented cable with 4000 kg (9000 lb) of gear that was moored to the sea floor under the ice (box fig. 4). The mooring, supported by submerged floats, carries conductivity-temperature recorders to monitor salinity and temperature changes, current meters to record speed and direction of flow, a current profiler to detail ice drift and vertical structure of currents, and sonar to measure ice thickness. Each instrument must record its data internally because a satellite cannot retrieve data from under the ice. The data were recovered in

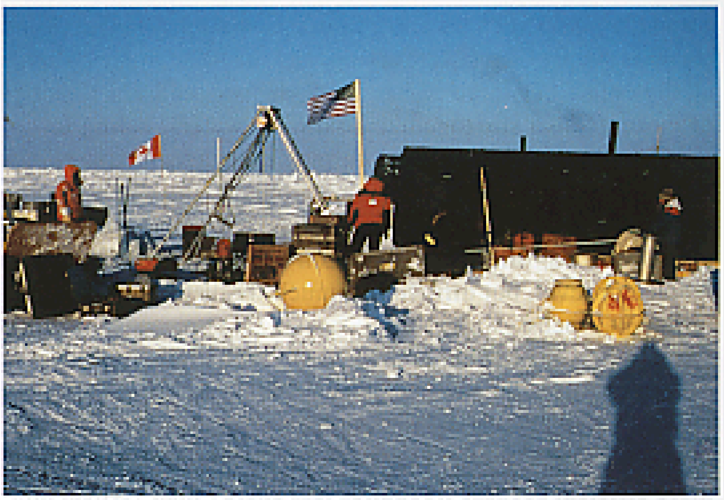
Box Figure 2 A polar bear is tranquilized and tested for contaminants during the 1994 Arctic Ocean Section expedition.

Box Figure 3 The Canadian Coast Guard icebreaker *Des Groseilliers* serves as the support base for SHEBA research huts on the Arctic ice.

the following year when the mooring was retrieved. Another part of the 2001 program includes using ski-equipped aircraft to establish five camps over a 485 km (300 mi) route from the pole to Alaska. At each camp (box fig. 5), members of the research team make measurements through the ice and collect samples for chemical



Box Figure 4 Deploying a 4250 m mooring at the North Pole. The mooring is outfitted with a suite of oceanographic instruments. An acoustic Doppler current profiler is located near the top of the mooring.



Box Figure 5 The building on the ice combines living quarters and a scientific laboratory. Equipment includes an A-frame and buoys ready for deployment through the ice.

tests. These data continue the effort of 2000, when similar data were obtained over a 565 km (350 mi) route between the pole and Canada. The data that are being collected will be combined with satellite and meteorological data to study changes in the Arctic Ocean and to determine how fast the changes are occurring.

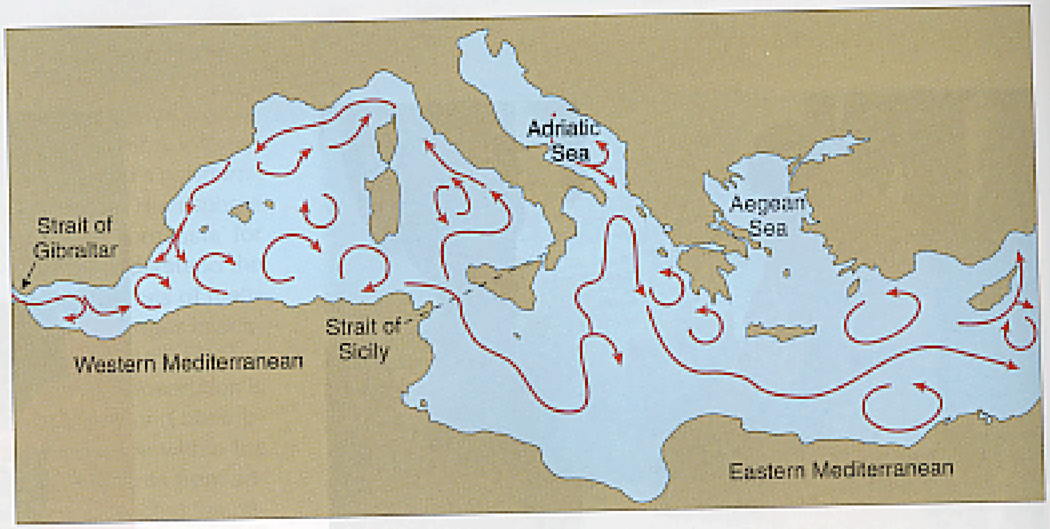


Figure 8.9 Surface currents in the Mediterranean Sea. A *line* through the Strait of Sicily separates the Mediterranean Sea into the western Mediterranean basin and the eastern Mediterranean basin. Deep and intermediate waters are formed in the eastern basin, move through the Strait of Sicily into the western basin, continue westward, and exit through the Strait of Gibraltar into the Atlantic Ocean. Before 1987, cool, high-salinity, deep water was formed principally in the Adriatic Sea; by the winter of 1995, the source of this water had changed to the Aegean Sea.

into the Mediterranean Sea to replace seawater that exits the Mediterranean at depth (fig. 8.9).

Oceanographers are interested in what happens in the Mediterranean Sea because the forces that drive the Mediterranean system are comparable to the larger-scale processes governing the open ocean; thus, the Mediterranean Sea can be used as a model for the larger ocean system.

Internal Mixing

Mixing between waters in the ocean is most active when turbulence and energy of motion are available to stir the waters and blend their properties. At the sea surface, wind-driven waves and currents supply energy for mixing, and the tides create currents at all depths. The large eddies that may form at the boundaries of currents also stir together dissimilar waters, acting to homogenize them. When surface currents coverage, mixing at current boundaries may produce caballing of the mixed water. When currents and their associated turbulence are weak, mixing is reduced. Mixing by diffusion occurs continually at the molecular level, but diffusion is much weaker than mixing by turbulent processes.

If a parcel of water is displaced vertically by turbulence, buoyancy forces tend to return the parcel to its original density level. Therefore, vertical mixing between the water types that form the oceans' internal layers is weak. Horizontal mixing is more efficient because it requires less energy than vertical mixing. A parcel of water displaced horizontally along a surface of constant density remains at its new position and shares its properties with the surrounding water.

In areas under warm, high-salinity surface water with an appreciable salinity and temperature decrease with depth, internal vertical mixing processes occur despite the stability of the water column. Vertical columnar flows, approximately 3 cm (5 in) in diameter, are called salt fingers; they develop and mix the water vertically, causing a stair-step salinity and temperature change with depth. This phenomenon is caused by the ability of seawater to gain or lose heat faster by conduction than it gains or loses salt by diffusion. This causes the density of the vertically moving water to change relative to that of the surrounding water, and the moving water is propelled either up or down. Salt fingers mix water over limited depths, creating homogeneous layers 30 m (100 ft) thick. These layers exist from about 150–700 m (500–2300 ft) deep and are estimated to occur over large areas of the oceans when the required conditions are present.

8.4 Measurement Techniques

Measuring the salinity, temperature, dissolved gases, nutrients, suspended matter, and other characteristics of seawater in situ requires that the oceanographer devise specialized sampling and measuring equipment and a platform from which the equipment can be used or deployed. The research vessel is the traditional platform from which samples are taken and measurements are made (see the photo essay "Going to Sea," pages 266–269). These vessels are equipped with winches and cables that launch, lower, and retrieve instruments; laboratories for specialized onboard research; and a large array of measuring devices, including depth recorders, sonar, speed and direction sensors, atmospheric and solar radiation sensors, and many more. However, research time at sea is expensive, for vessels require fuel, living amenities, and professional merchant crews as well as oceanographers. Total vessel and crew costs are high: $25,000 and more per operating day, not including the scientific party and equipment. Therefore, researchers need to be able to gather accurate information in the least amount of time.

The oceanography of the first fifty years of the twentieth century depended on robust mechanical devices that were lowered into the sea, took their samples at preset depths, and returned those samples to a ship for later analysis, but their use is decreasing rapidly.

Although early water bottle data were sparse, they were sufficient to indicate the depth structure of the oceans. Estimated flow rates of water in deep-ocean circulation were made by combining dissolved oxygen measurements and estimated average values of oxygen utilization. Using these values, researchers calculated the time required for a water parcel to leave the ocean surface and arrive at depth. The calculated times coupled with water parcel positions yielded water transport rates.

Direct measurements of transport rates were later made by suspending neutrally buoyant floats in the moving water. The floats emitted periodic sound pulses (pings) that were used to locate them in the water flow. Indirect measurements of ocean circulation also involved isotope tracers that change their properties with time. Tritium, a radioactive isotope resulting from nuclear testing in the 1950s and 1960s, has a half-life of 12.45 years. It was used in large ocean-circulation studies concerned with short-term displacements of water at depth. Longer-term changes in the circulation used isotopes such as C^{14}, which has a much longer half-life, 5570 years.

The conductivity-temperature-depth sensor, or CTD (fig. 8.10), is today's workhorse for determining water properties and providing a depth profile of salinity by measuring electrical conductivity of the seawater. A temperature profile is obtained with an electrical resistance thermometer, and depth readings are made with a pressure sensor. The CTD is lowered in a protective cage that contains a series of water bottles and may also carry other sensors such as pH probes, chlorophyll sensors, optical scanners, and dissolved oxygen sensors. Conductivity and temperature are monitored continuously as the CTD is lowered, and the data are returned to the ship as electronic signals through the suspending cable. Onboard ship, the CTD data may be fed directly into a computer, recorded on a chart, or made available as numerical data. Continuous profiles of data to several thousand meters can be acquired in this way, whereas the old water bottles collected only one sample at each depth for later analysis.

CTD systems still require a research vessel to stop and remain on station while the instrument is lowered, measurements are made, and the CTD is recovered. However, the CTD can also be left in place, suspended from a surface vessel or buoy; the collected data can be stored in digital form for retrieval at another time.

Seasoar (fig. 8.11) is a modified CTD device that allows sampling while the vessel is underway. The Seasoar is a winged vehicle that is towed behind a ship. The device controls its depth by diving planes and can be compared to a kite flying upside down underwater. Its vertical range extends from the surface to about 350 m (1100 ft), and its dive cycle takes about ten minutes to complete, compared with the standard CTD, which, with stop and start time for the vessel, takes about forty minutes. The Seasoar's sensors record continuously, both horizontally and vertically, while it dives.

What is required to sample an ocean and determine its structure and circulation as well as any changes that may be occurring? In the 1980s, most oceanographers would have said more time, more effort, more equipment, and more money than are available. But in the 1990s, a program was developed to do just that. The very large and very successful World Ocean Circulation Experiment (WOCE), a multi-institutional, multifaceted program, combined satellite data, buoys, direct measurement of currents and water properties, and the Global Positioning System. Data from this program greatly improved oceanographers' abilities to model the oceans and understand ocean-atmosphere interactions.

Figure 8.10 Retrieving a conductivity-temperature-depth sensor, or CTD. The CTD is attached below a rosette of water bottles. Data are relayed to the ship's electronic processing and data center. Water samples are taken when an interesting water structure is found or when water samples are required to calibrate the CTD.

In the spring of 2003, the National Oceanographic and Atmospheric Administration (NOAA) sponsored a cruise from Iceland to Madeira in the Atlantic Ocean and then on to Fortaleza, Brazil. This two-month trip repeated one of the sections of the WOCE program, allowing researchers to look for patterns of natural events that help to explain observed changes in oceanic structure and circulation since the WOCE cruise. Researchers also looked for patterns of natural events to guide them in planning future cruises.

To sample larger ocean areas than can be served by a single research ship and for continuous sampling over long periods of time, large, instrumented buoys are moored at sea and their data transferred by satellite or radio link to a research ship or laboratory for analysis. Other surface buoy systems are released to drift with the currents, monitoring water properties at changing locations and reporting back to ship or shore via satellite. Small buoys also drift with the surface currents, sink

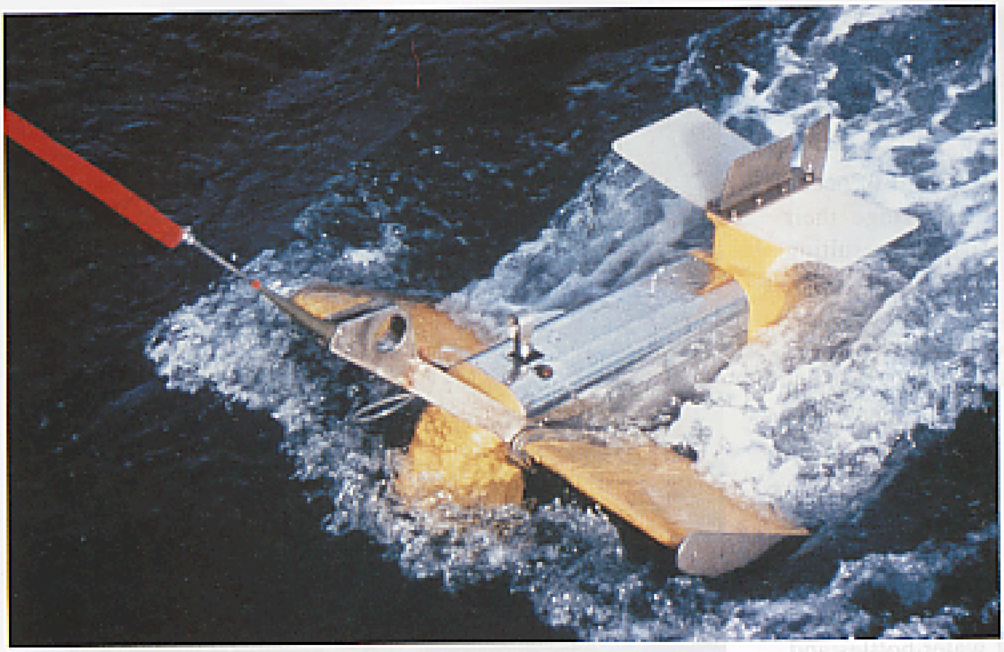


Figure 8.11 Seasoar is a conductivity-temperature-depth sensor that is towed by a ship. The wings allow Seasoar to move vertically while it is towed horizontally.

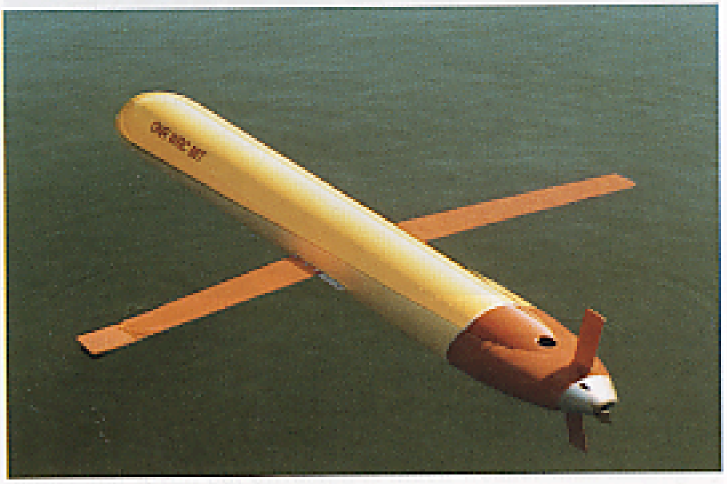


Figure 8.12 The SLOCUM glider has multiple sensors designed to monitor water properties independently for up to five years. While at the surface it is able to send data and update its navigation by satellite.

to predetermined depths, and then return to the surface to drift again. These instruments monitor water properties as they rise to the surface and send this information back to a research vessel or a satellite.

Continual improvements in electronics, computers, and sensor systems have resulted in oceanographers devising independent and relatively inexpensive devices that are placed in the sea in large numbers. These independent, untethered instruments sink to predetermined depths, making measurements as they rise through the water column on their return to the surface. These data-gathering devices are known as profilers and carry a wide variety of sensors. The availability of these tools is beginning to provide the density of data needed to determine both long- and short-term trends in the oceans.

One type of profiler functions as a winged glider (fig. 8.12; see the box titled "Ocean Gliders"). When this profiler sinks and rises, it flies down and up at an angle, measuring properties along a diagonal over its operating depth. This flying ability allows the profiler to wander partially independent of the currents. The Argo profiler (see fig. 1.19) does not fly but sinks and rises vertically only. Lateral displacement is caused by the currents encountered over its vertical path. Once at the surface, both types of profilers broadcast their recorded data and positions to a satellite, then sink and repeat the process.

Specialized satellites measure a large variety of sea surface and atmospheric conditions, sea topography, wind speeds, plankton abundance, air and water temperatures, waves, and more. Satellites are very expensive monitoring devices, but their ability to survey the global oceans repeatedly and collect huge amounts of data makes them cost-effective as they collect data that cannot be obtained by any other means.

Not all marine studies require a large research vessel with its sophisticated and expensive equipment. Inshore, shallow-water studies usually use smaller vessels with less-sophisticated equipment. Small winches driven by power or even by hand are used to lower mechanical water bottles with their thermometers to the required depth. Lightweight electronic instruments are also available and include devices to measure conductivity, temperature, dissolved oxygen, optical properties, pH, and fluorescence of chlorophyll.

8.5 Practical Considerations: Ocean Thermal Energy Conversion

An indirect form of solar energy and an alternative to fossil fuel production of energy is found in the sea. The transparency and heat capacity of water allow large amounts of solar energy to be stored in the ocean, and this heat can be extracted independent of daily and seasonal changes in the available solar radiation.

Ocean thermal energy conversion (OTEC) depends on the difference in temperature between surface water and water at 600–1000 m (2000–3300 ft) depth. There are two types of OTEC systems: (1) closed cycle (fig. 8.13*a*), which uses a contained working fluid with a low boiling point, such as ammonia, and (2) open cycle (fig. 8.13*b*), which directly converts seawater to steam.

In a closed system, the warm surface water is passed over the evaporator chamber containing the ammonia, and the ammonia is vaporized by the heat derived from the warm seawater. The vapor builds up pressure in a closed system, and this gas under pressure is used to spin a turbine, which generates power. After the pressure has been released, the ammonia

An oceanographic research vessel can monitor only a small portion of an ocean during any one cruise, and the cost is very high (see the photo essay "Going to Sea"). To increase the size of the area that can be surveyed, as well as lower the cost, oceanographers have been experimenting with independent, unstaffed vehicles (see figs. 8.13 and 8.14). A new type of survey vehicle is an ocean glider, which functions in the water much as a conventional sailplane does in the atmosphere. Unlike a sailplane, which has to be towed aloft and ultimately falls back to the ground, an ocean glider has the ability to change its buoyancy so that it can glide up as well as down.

An oceanographic glider, like an airborne glider, has no propeller but is powered by its buoyancy. Buoyancy is changed by adjusting glider volume in order to sink or rise. The presence of wings translates this vertical force into motion along a slanting glide path. The limited onboard battery energy is devoted to buoyancy control, sampling sensors, navigation, and communication systems. Through energy efficiency, the glider remains operational for long periods, up to a year, and collects and reports large quantities of data.

At present, research groups at the Scripps Institution of Oceanography, the Woods Hole Oceanographic Institution, the Webb Research Corporation, and the University of Washington are developing gliders with the support of the U.S. Navy, the U.S. National Science Foundation, and the National Oceanic and Atmospheric Administration. Different glider designs are being used to address different ocean sampling needs, such as performing long open-ocean transects, maintaining geographic position while profiling vertically, or operating in shallow regions. One under development by Webb Research, known as the SLOCUM Glider, extracts energy from ocean thermal stratification to change buoyancy (see fig. 8.12). Others use battery energy for propulsion.

At the University of Washington, Seagliders are being developed and evaluated (box fig. 1). Seagliders are small (1.8 m [7 ft] long), light (52 kg [115 lb]) battery-powered gliders that can be easily launched and retrieved from small vessels rather than relying on research ships. They alternatively dive and climb through the water column as oil is transferred between an internal reservoir and an external bladder. When the glider is launched at the surface, its density is just slightly less than the density of seawater. A small amount of oil is bled from the external bladder to the inside of the pressure case, decreasing the bladder volume and increasing the density of the Seaglider, causing it to sink. When sinking motion starts, the wings on the Seaglider provide lift, and the glider moves forward as it sinks. When the vehicle reaches its programmed depth, an electric pump transfers oil from inside the pressure case to the external bladder, decreasing the vehicle density. As the less-dense glider rises toward the surface, the wings act to hold it down so that, it moves forward. Its forward speed is only about 0.25 m/s (0.5 knots), but because drag increases quadratically with speed, the glider has a range and endurance much greater than conventional propeller-driven autonomous underwater vehicles.

Box Figure 1 A Seaglider being launched for a trial ocean run.

Seagliders carry sensors and recorders for temperature, salinity, dissolved oxygen, fluorescence, and optical backscatter. GPS navigation and telemetry equipment are also onboard; antennae are housed in the trailing wand (box fig. 2). At the surface, the glider determines its position using GPS and sends data via a global satellite phone link. Researchers are able to communicate with the glider and guide it under remote control.

To measure surface currents, the drifting glider can be tracked by GPS. When the glider surfaces from a dive, the difference between the glider's actual position and its intended position is the result of displacement by the average of the currents over both its sinking and rising paths.

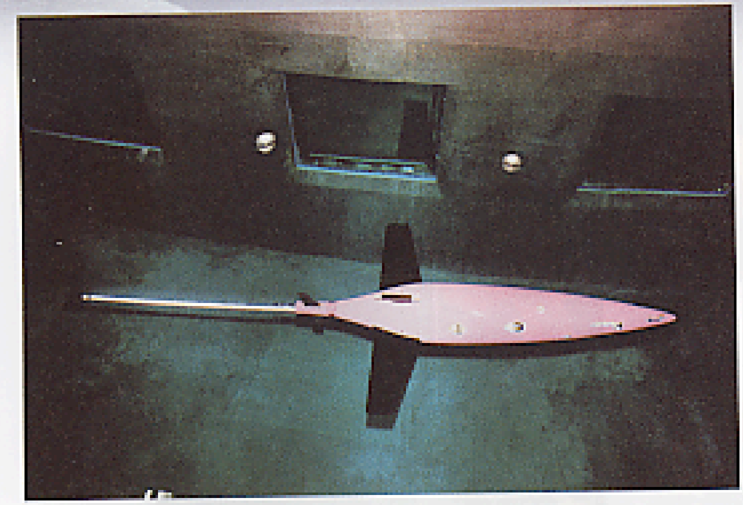
Seagliders are designed to dive as deep as 1000 m (3000 ft) and cover as much as 10 km (5.3 nautical mi) horizontally in one dive cycle. In a six-month mission, they can travel as much as 5000 km (2700 nautical mi) through the ocean.

(continued)

is passed to a condenser, where it is cooled by cold water pumped up from depth. Cooling condenses the ammonia to the liquid state, and the liquid is pumped back to the evaporator to repeat the cycle.

In an open system, large quantities of warm seawater are converted to steam in a low-pressure vacuum chamber, and the steam is used as the working fluid. Because less than 0.5% of the incoming water is turned into steam, large quantities of warm water must be used. The steam passes through a turbine and into a condenser cooled by cold water from depth and condenses to desalinated water.

OTEC plants using either system can be located onshore, offshore, or on a ship that moves from place to place. Figure 8.13c is an engineering concept design for an

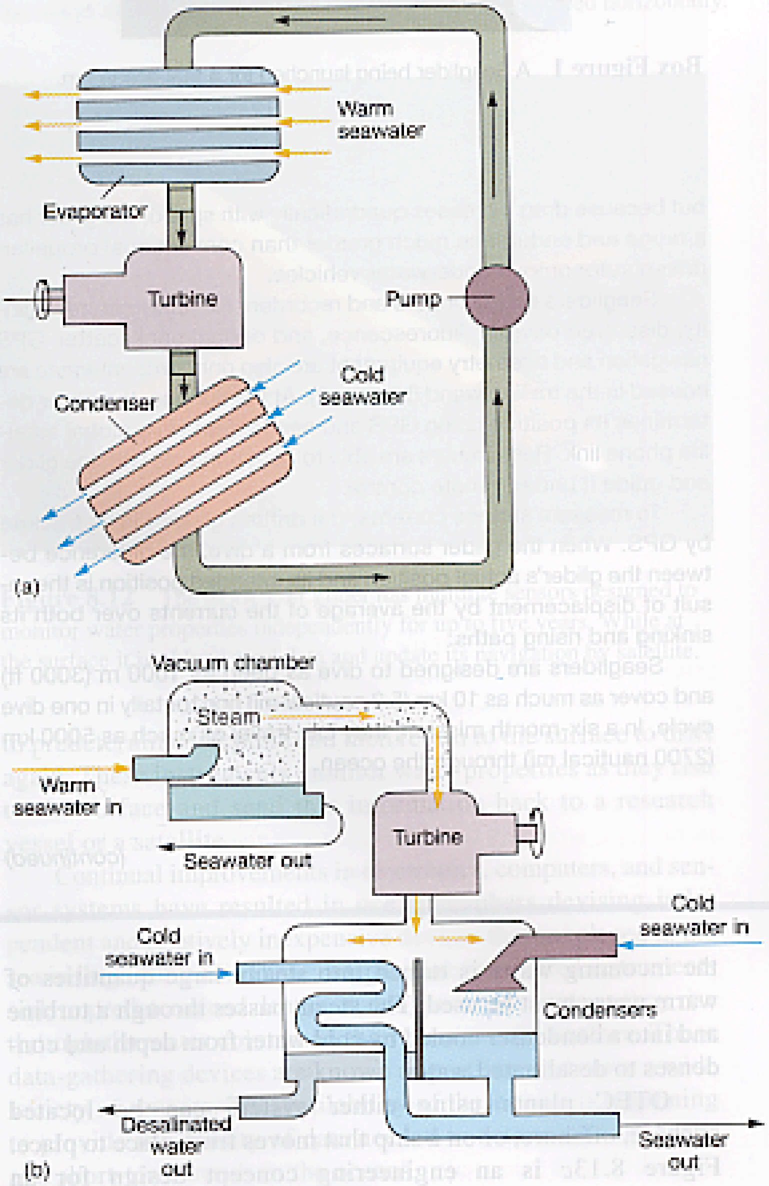


Box Figure 2 A Seaglider moving upward in a testing tank.

While building a glider costs roughly as much as buying an expensive automobile, its operational cost is quite low. A single, one-year mission costs about the same as one day of research vessel operation. The production and annual operational cost of gliders will decrease when they are produced in quantity. Even in their development stage, gliders have been able to collect observations that are prohibitively costly using ship-based means.

To Learn More About Ocean Gliders

Eriksen, C. C., T. J. Osse, R. D. Light, T. Wen, T. W. Lehman, P. L. Sabin, J. W. Ballard, and A. M. Chiodi. 2001. Seaglider: A Long Range Autonomous Underwater Vehicle for Oceanographic Research. *I.E.E.E. Journal of Oceanic Engineering,* 2001 26(4):421–23.



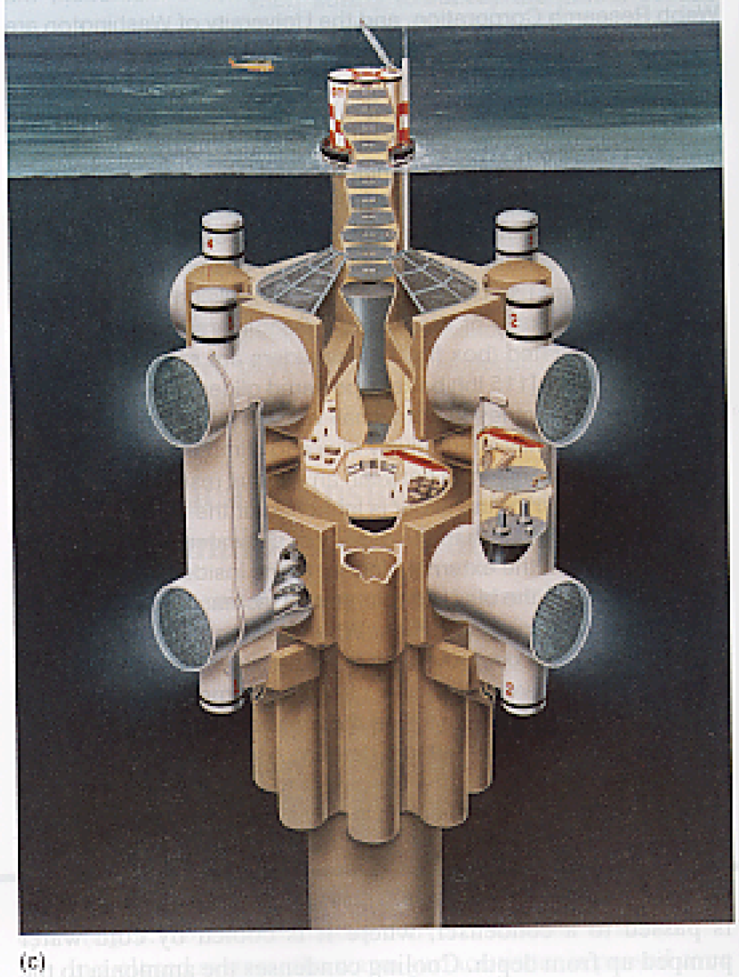


Figure 8.13 (a) The simplified working system of an ocean thermal energy conversion (OTEC) closed-system electrical generator. (b) The simplified working system of an OTEC open-system electrical generator. (c) The concept drawing for the Lockheed OTEC Spar.

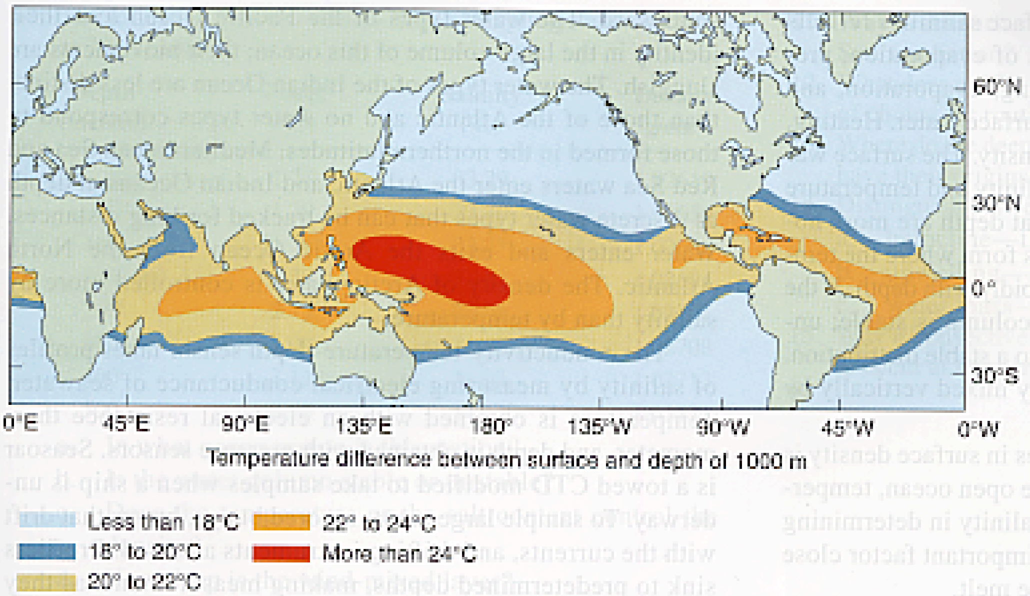


Figure 8.14 OTEC plants are ideally suited to areas with a large temperature difference between the surface and deep waters (at least 20°C). The conditions are generally found in tropical and equatorial latitudes. Image from the National Renewable Energy Laboratory.

independent, free-floating OTEC plant. The Lockheed Spar has not been built, and there are no present plans to do so. OTEC requires at least a 20°C difference in temperature between surface and depth to generate useful amounts of energy. These power plants would be located at latitudes between approximately 25°N and 25°S where both warm surface water and cold deep water are available; seawater temperature difference over depth must average 22°C (fig. 8.14). W. H. Avery and C. Wu (*Renewable Energy from the Ocean, a Guide to OTEC*, Oxford University Press, New York, 1994) estimated that the area of the world's oceans meeting OTEC requirements is 60×10^6 km^2 (23×10^6 mi^2). The total power generated within this area would exceed 10×10^6 megawatts (MW); the total U.S. electricity generating capacity is about 1.7×10^6 MW.

Engineers studying the potential and feasibility of OTEC have estimated that about 0.2 MW of usable power can be extracted per square kilometer of tropical ocean surface; that is about 0.07% of the average absorbed solar energy. The process is considered safe and environmentally benign; however, the pumping of cold water from depth to the sea surface brings nutrients from depth and liberates them into the nutrient-poor tropical surface waters, changing the productivity of the surface water.

The Natural Energy Laboratory of Hawaii near Keahole Point operated a land-based, open-system OTEC plant for six years (1993–98). The project produced a net of 100 kW of power per day using the temperature difference between the water at the ocean surface and the water about 800 m (2500 ft) below the surface. The plant also produced 26,000 l (7000 gal) of desalinated water per day. Aquaculture projects using the nutrient-rich water from depth included production of oysters, shrimp, fish, black pearls, and various seaweeds.

Work on OTEC technology slowed in the late 1990s because it was too costly compared to generating energy using relatively inexpensive oil. The dramatic increase in the price of oil in recent years has resulted in a renewed interest in OTEC. In 2006 plans for building the world's two largest OTEC power plants in Hawaii were announced. One power plant is planned to be operational at the Natural Energy Laboratory of Hawaii Authority in Kona in 2008. This power plant will produce up to 1.2 megawatts, of which about one-third will be used to run pumps and keep the system going. That leaves a net production of 800 kilowatts. The second power plant is planned to be located at an undisclosed ocean location for military use. This plant will produce a net of 8.2 megawatts plus 1.25 million gallons of fresh water a day.

Other ideas for using cold deep-ocean water, or DOW, pumped from 630 m (2000 ft) include air conditioning and industrial cooling systems. The condensate that forms on pipes carrying DOW is fresh water, formed at a rate of about 5% of the flow of the cold water. A flow of 76,000 l (20,000 gal) per minute generates an estimated 3800 l (1000 gal) per minute of fresh water. Lobsters, flat fish, shrimp, abalone, oyster, and other organisms have been raised in the tropics using DOW. Tropical agricultural experiments have also used DOW, pumping it through pipes placed at root depth in the soil. This system chills the soil to 10°C and produces freshwater condensate on the pipes and on the soil.

Summary

The absorption and exchange of energy at the sea surface govern the properties of surface seawater. Sea-surface exchanges of heat, radiant energy, and water alter the temperature and salinity of the surface water and affect the density of the water. Many combinations of salinity and temperature can produce seawater of the same density. When waters with different properties but the same density are mixed, the resulting water has a greater density than either of its components. The density-driven vertical circulation that results is known as caballing. The waters of different densities produced at the sea surface and the resulting vertical circulation create a layered ocean that is primarily stably stratified.

The geographical distribution of surface salinities reflects Earth's latitudinal and seasonal patterns of evaporation, precipitation, and sea-ice formation. Cooling, evaporation, and freezing increase the density of the sea-surface water. Heating, precipitation, and ice melt decrease its density. The surface water changes its density with changes in salinity and temperature that are keyed to latitude. The densities at depth are more homogeneous. Thermoclines and haloclines form where the temperature and salt concentrations change rapidly with depth. If the density increases with depth, the water column is stable; unstable water columns overturn and return to a stable distribution. Neutrally stable water columns are easily mixed vertically by winds and waves.

Vertical circulation driven by changes in surface density is known as thermohaline circulation. In the open ocean, temperature is generally more important than salinity in determining the surface density. Salinity is the more important factor close to shore and in areas of large seasonal ice melt.

Water sinks at downwellings and rises at upwellings. A downwelling occurs at the convergence of surface currents and transfers oxygen to depth. Upwellings bring nutrients to the surface and occur at zones of surface current divergence.

The oceans are layered systems. The layers (or water types) are identified by specific ranges of temperature and salinity. The water types of the Atlantic Ocean are formed at the surface at different latitudes. They sink and flow northward or southward. The water types of the Pacific Ocean lose their identity in the large volume of this ocean; their movements are sluggish. The water types of the Indian Ocean are less distinct than those of the Atlantic and no water types correspond to those formed in the northern latitudes. Mediterranean Sea and Red Sea waters enter the Atlantic and Indian Oceans at depth as discrete water types that can be tracked for long distances. Water enters and exits the Arctic Ocean from the North Atlantic. The density of Arctic water is controlled more by salinity than by temperature.

The conductivity-temperature-depth sensor takes profiles of salinity by measuring electrical conductance of seawater. Temperature is obtained with an electrical resistance thermometer, and depth is obtained with pressure sensors. Seasoar is a towed CTD modified to take samples when a ship is underway. To sample large areas, moored buoys, buoys that drift with the currents, and drifting instruments are used. Profilers sink to predetermined depths, making measurements as they rise. All are able to send their positions and data via satellite. Satellites also monitor the oceans directly. Shallow-water measurements can be made with mechanical water bottles and thermometers.

Ocean thermal energy conversion generates energy by using the difference in temperature between ocean surface water and deeper water. A land-based, open-system OTEC plant is operated in Hawaii.

Key Terms

All key terms from this chapter can be viewed by term or definition when studied as flashcards on this book's website at www.mhhe.com/sverdrup10e.

caballing, 201
pycnocline, 202
thermocline, 202
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stable water column, 202
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Mediterranean Intermediate Water (MIW), 207
ocean thermal energy conversion (OTEC), 212

Study Questions

1. Why does the mixing of two water types at the sea surface result in a mixed water type that sinks to a deeper depth?
2. What natural processes alter the surface salinity of the oceans? How do these processes vary with latitude?
3. Describe the changes in water density in the upper ocean layer over the annual cycles at tropical, polar, and temperate latitudes. Indicate when periods of stable and unstable conditions exist in the upper water column.
4. Why are water layers more prominent in the Atlantic Ocean than in the Pacific Ocean?
5. How do salt fingers form, and how do they contribute to vertical mixing?
6. How can heat be transferred from place to place in the oceans? Why is this heat transfer ignored when considering the world's total heat budget?
7. If the upper layers of an ocean area are homogeneous in salinity, explain why the thermocline coincides with the pycnocline.
8. How does Mediterranean Sea water alter the salinity and temperature of the North Atlantic? Use figures 8.6 and 8.7*a* and *b*. At what depth does this change occur?
9. A water sample taken at 4000 m in the Atlantic Ocean has a salinity of 34.8‰ and a temperature of 3°C. At approximately what latitude was this water last at the surface?