Lecture Notes

INTRODUCTION TO THE OCEANS

…

TECTONICS

PHYSICS AND CHEMISTRY

MEASURING TECHNIQUES AND INSTRUMENTS

…

Table of Contents

Ocean-Floor Bathymetry

The term bathymetry is defined as the depth of water relative to sea level. Thus bathymetric measurements can determine the topography of the ocean floor, and have shown that the sea floor is varied, complex, and ever-changing, containing plains, canyons, active and extinct volcanoes, mountain ranges, and hot springs. Some features, such as mid-ocean ridges (where oceanic crust is constantly produced) and subduction zones, also called deep-sea trenches (where it is constantly destroyed), are unique to the ocean floor. [*]

Bathymetric mapping involves the production of ocean and sea maps based upon bathymetric data (see historic map). Bathymetric maps represent the ocean depth as a function of geographical coordinates in the same way topographic maps represent the altitude of Earth's surface at different geographic points. The most popular type of bathymetric maps are ones on which lines of equal depths (called isobaths) are represented.

Bathymetric Techniques

For hundreds of years, the only way to measure ocean depth was the sounding line, a weighted rope or wire that was lowered overboard until it touched the ocean floor. Not only was this method time-consuming, it was inaccurate; ship drift or water currents could drag the line off at an angle, which would exaggerate the depth reading. It was also difficult to tell when the sounding line had actually touched bottom.

England's Sir John Murray compiled this bathymetric (depth) chart of the North Atlantic in 1911. Murray's chart went far beyond American naval officer Matthew Maury's first attempt at bathymetric mapping in 1855. In addition, Murray's map gave birth to the idea of the Telegraphic Plateau, a submarine land formation from Canada to the British Isles, across which the first transatlantic cable was laid.

In the twentieth century, sounding lines were entirely replaced by sonar systems. Sonar (*so* und *na* vigation *r* anging), invented during World War II (1939–1945) measures distances by emitting a short pulse of high-frequency sound and measuring the time until an echo is heard. After the

war, ships with sonar units attached to their hulls crisscrossed the world's oceans systematically, measuring depth. The data collected made possible complete bathymetric maps of the world's oceans. For the first time, scientists knew what 70 percent of Earth's surface really looked like (radar, which produces images by bouncing radio waves rather than sound waves off distant objects, cannot be used for bathymetry because water absorbs radio waves).

Many sonar techniques have been developed for bathymetry. When high-resolution images are desired, an underwater unit may be towed behind a ship, scanning to the left and right with multiple sonar beams (sidescan sonar). [*] Furthermore, orbiting visible-light cameras image the bottoms of some shallow waters, while satellite radar maps deep-sea topography by detecting the subtle variations in sea level caused by the gravitational pull of undersea mountains, ridges, and other masses.

The Ocean Floor in Cross-Section

The oceans begins, of course, at the shore, the irregular boundary where the surface of a continent descends first to sea level and then beneath it. If the depth of the ocean is measured along a line drawn straight out from a continental shore, the following sequence of bottom features are typically seen.

Continental Shelf.

For many miles out, the ocean is only a few hundred feet deep and gets deeper quite slowly (i.e., slopes at an angle of 0.1°, or 1.7 meters per kilometer [9 feet per mile]). This flat, wide margin is found around every continent and is known as the continental shelf. The average width of a continental shelf is 70 kilometers (43 miles).

Continental Slope and Rise.

The continental shelf ends at a sudden dropoff called the shelf break. Beyond the shelf break, the slope of the ocean floor becomes much steeper, typically a 4° slope, or 70 meters per kilometer (370 feet per mile). This steep embankment is called the continental slope and is grooved by submarine canyons and gullies. The continental slope is about 16 kilometers (10 miles) wide, on average, and descends to a depth of about 2.4 kilometers (1.5 miles). There it ends as the slope moderates to a mere degree or two from horizontal. This gradual zone, which may be several hundred miles wide, is called the continental rise. It is composed of fine-grained continental sediments (silt and clay) washed down the many submarine canyons that notch the continental slope.

Abyssal Plain.

The abyssal plain, which is the deepest, most level part of the ocean, is found where the continental rise ends, at a depth of about 4 kilometers (2.5 miles). The abyssal plain is dotted with thousands of small, extinct volcanoes called abyssal hills. The abyssal plains of the Atlantic appear smooth because its abyssal hills are buried under a thick blanket of continental sediment, but in the Pacific Ocean basin, which is ringed by trenches that trap sediments before they can spread over the ocean floor, tens of thousands of unburied abyssal hills have been observed. Abyssal hills more than 1 kilometer (0.6 mile) high are called seamounts, and seamounts with flat tops are called guyots or tablemounts. Guyots are drowned volcanic islands that become submerged due to subsidence of the oceanic **lithosphere** .

Mid-Ocean Ridge.

Beyond the abyssal plain, which may be several hundred kilometers wide, the ocean floor begins to ascend again with a gentle slope. This area is the flank of the mid-ocean ridge, a long, undersea mountain chain that usually extends down the middle of the ocean. The Mid-Atlantic Ridge, for example, snakes down the middle of the Atlantic most of the way from the North Pole to Antarctica. [*] As the ocean floor climbs slowly toward the center of the mid-ocean ridge its sediment blanket gets thinner and its surface more irregular. Here, the ocean floor is marked by thousand-mile cracks called fracture zones that lie at right angles across the mid-ocean ridge.

Rift Valley.

Along the center of the mid-ocean ridge is the rift valley, a deep V-shaped notch. From this valley, new oceanic crust is constantly being extruded from Earth's **mantle** by processes not yet fully understood. Twin sheets of fresh, mile-thick crust emerge from the mantle along the sides of the rift valley and flow slowly away from it in opposite directions. In the case of the Mid-Atlantic rift valley, one sheet flows east and the other west, each moving at about half an inch per year. The older, more distant parts of these growing sheets of crust are gradually covered by sediments, eventually becoming the abyssal plains and continental rises described above. These growing sheets of crust have the effect of forcing the distant continents farther apart, a process called sea-floor spreading.

Subduction Zones.

Beyond the rift valley, the depth sequence described above is observed in reverse: a gently declining slope of ridges and fractures, an abyssal plain, a continental rise, a steep continental slope, a somewhat flat continental shelf, and finally dry land again. This would be the case in crossing the Atlantic Ocean. Elsewhere, however, such neat symmetry is not always found. Some continental margins are characterized by subduction zones, whereby oceanic crust is destroyed beneath the edge of a continental plate.
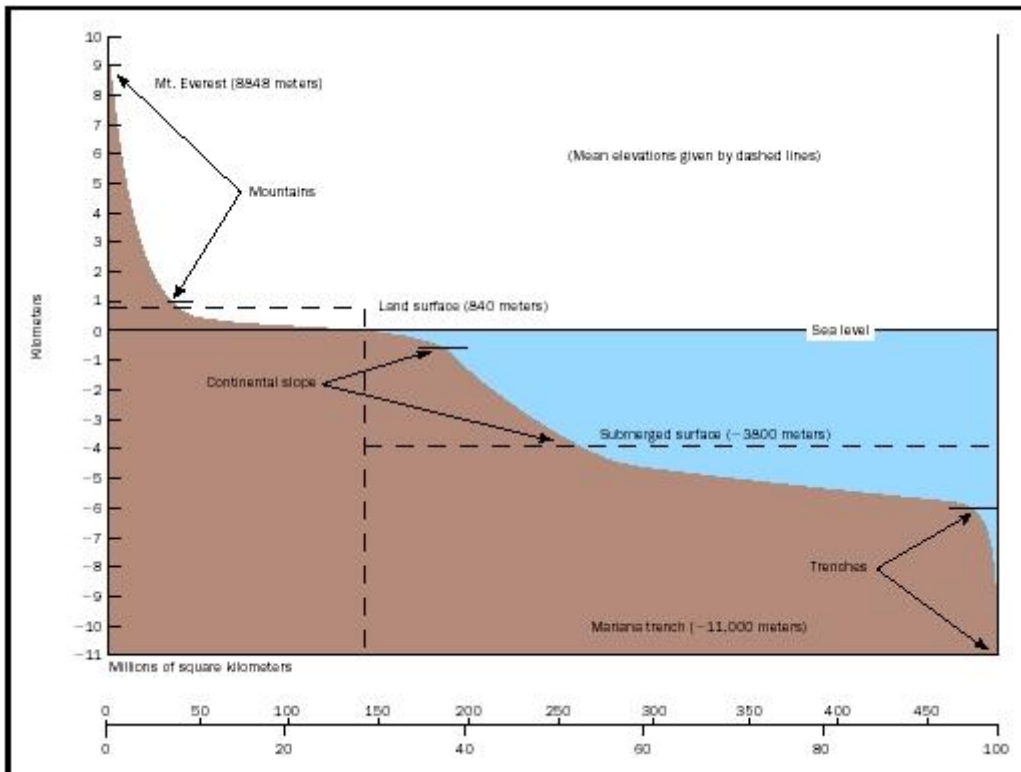
The hypsographic curve shows the amount of Earth's surface at various elevations and depths. Horizontal dashed lines indicate average height of the continents at 840 meters (2,750 feet) above sea level, and average depth of the oceans at 3,800 meters (12,460 feet) below sea level. The vertical dashed line marks the division between land and sea at present-day sea level.

The disappearance of oceanic crust into subduction zones is occurring primarily along the western, northern, and eastern edges of the Pacific Ocean. These subduction zones are the deepest places in the ocean, varying from 7 to 11 kilometers (4 miles to 6.8 miles) in depth. Along the length of a typical subduction zone or deep-sea trench, a sheet of old oceanic crust is forced beneath continental crust. The oceanic crust that is forced to submerge or subduct in this manner slides down into the mantle at an angle.

The Hypsographic Curve

The elevation features of the ocean floor—shelves, slopes, rises, plains, ridges, and trenches—are quantitatively summarized by oceanographers, along with the distribution of dry land at different altitudes, in a graph called the hypsographic curve (see the figure). The hypsographic curve shows what percentage of Earth's surface rises above present-day sea level to a given height, or sinks below it to a given depth. The curve shows that a small percentage of Earth's surface consists of high-altitude continental mountains, 30 percent of the surface consists of continental lowlands. Continental shelves and slopes account for about another 15 percent, abyssal plains and mid-ocean ridge systems for almost 50 percent, and deep-sea trenches account for a small fraction. The fact that Earth's surface is comprised mostly of flat plates of continental and oceanic crust, with mountains and deep-sea trenches occurring only along their edges where the plates collide, gives the hypsographic curve its characteristic shape.

Mid-Ocean Ridges

The mid-ocean ridge is an interconnected system of undersea volcanoes that meander over the Earth like the raised seams on a baseball. It is a continuous 40,000-mile (60,000-kilometer) seam that encircles Earth and bisects its oceans. The mid-ocean ridge represents an area where, in accordance with **plate tectonic** theory, **lithospheric plates** (also called tectonic plates) move apart and new crust is created by magma (molten rock) pushing up from the **mantle** . The mid-ocean ridge system is an example of a divergent (rather than a convergent or transform) plate boundary.

The mid-ocean ridge system has been understood only since the development and acceptance of plate tectonic theory in the 1960s. Four major scientific developments spurred the formation of the theory: (1) demonstration of the young age of the ocean floor; (2) confirmation of repeated reversals of Earth's magnetic field in the geologic past; (3) emergence of the seafloor-spreading hypothesis and associated recycling of the oceanic crust; and (4) precise documentation that Earth's earthquake and volcanic activity is concentrated along subduction zones and mid-ocean ridges.

Ridge Characteristics

Mid-ocean ridges have different shapes (morphology) depending on how fast they are spreading, how active they are magmatically and volcanically, and how much tectonic stretching and faulting is taking place. Scientists believe that the most likely reason for the different morphologies is due to the strength of the ocean crust at these different sites, and how cold and brittle the upper part of the tectonic plate is.

Ridge Types.

There are two types of mid-ocean ridges: fast-spreading and slow-spreading. Fast-spreading ridges like the northern and southern East

A composite map showing the topography (relief) of the ocean floor clearly reveals the mid-ocean ridge system, which appears here as dark "seams" extending through the oceans. In the Atlantic Ocean, the mid-ocean ridge is called the Mid-Atlantic Ridge. It curves from the Arctic Ocean southward, through Iceland, down the center of the Atlantic, and around the bottom of Africa. There it divides, with one branch running up the Red Sea and the other going around Australia into the Pacific and north back to the Arctic. A branch off South America is called the East Pacific Rise.

Pacific Rise have smoother topography at the ridge crest, and look somewhat like domes. They have relief of 100 to 200 meters (328 to 656 feet). The East Pacific Rise moves at an average of 15 centimeters (5.9 inches) per year.

Slow-spreading ridges like the Mid-Atlantic Ridge have large, wide, rift valleys, sometimes as wide as 10 to 20 kilometers (6 to 12 miles) and very rugged terrain at the ridge crest that can have relief of up to 3.2 kilometers (2 miles). The Mid-Atlantic Ridge moves at an average of 2.5 centimeters (1 inch) per year.

Fast-spreading ridges are "hotter," meaning that more magma is present beneath the ridge axis, and that more volcanic eruptions occur. Because the plate under the ridge crest is hotter, scientists think that the plate responds to the divergent spreading process more fluidly, and that the ridge behaves like hot taffy being pulled apart. In this scenario, the ridge crest does not have a chance to subside (sink or settle).

At slower spreading ridges, the seafloor behaves more like a cold chocolate bar—when pulled, it cracks and breaks to form ridges and valleys. As the sheets of oceanic crust move away from the mid-ocean ridge, the rock is cooled and thus becomes heavier. After about 200 million years, the cooled lithospheric plate has become heavier than the **asthenosphere** that it rides over, and it sinks, thereby producing a subduction zone.

Fracture Zones.

Mid-ocean ridges do not form straight lines but are instead offset in many places by fracture zones, or transform faults. Fracture zones are thought to occur due to zones of weakness in the pre-existing continent before it was rifted apart. Most mid-ocean ridges are divided into hundreds of segments by fracture zones. Along the Mid-Atlantic Ridge, fracture zones occur at an average interval of 55 kilometers (34 miles). As the Mid-Atlantic Ridge is some 16,000 kilometers (10,000 miles) long, it is divided by fracture zones into about 300 distinct segments. The ridge crest and its associated faults are the locus of nearly all shallow earthquakes occurring in mid-ocean areas.

Water and Minerals.

Ocean water is constantly percolating through fissures (cracks) at the mid-ocean ridge. Downward-convecting cold ocean water meets the hot new crust far below the surface, and many types of metals such as sulfur, copper, zinc, gold, and iron are transferred to the water. This hot, mineral-laden water gushes back up through the cracks, forming hydrothermal vents. As the hot water, which can reach temperatures of 371°C (700°F), escapes from the vents and comes in contact with the near-freezing water of the ocean bottom, the metals quickly **precipitate** out of solution. The results are surging black clouds of particle-rich water called black smokers, which often erupt out of tall chimneys of previously deposited solidified mineral. [*] Because so much metal is spewed out, hydrothermal vents have been responsible for many of the world's richest ore deposits. These unique features also are found to harbor a diverse array of deep-ocean life.

Plate Tectonics

Plate tectonics is the unifying theory of geology that describes and explains that all earthquakes, volcanic activity, and mountain-building processes are caused by the gradual movement of rigid slabs of rock, called plates, that make up the Earth's surface layer. Given the expanse of geologic time, even modest movements—measured in centimeters or inches per year—result in substantial changes in the distribution of lands and oceans over millions of years.

Earth Structure

The Earth's internal structure can be viewed in two ways: either in terms of compositional layers, or in terms of layers of varying strength. There are three main compositional layers: the crust, mantle, and core. The crust, the outermost layer, is relatively buoyant and very thin compared to the mantle and core.

Beneath the oceans, the oceanic crust varies very little in thickness, generally extending only about 5 kilometers (3.1 miles), and is composed of **basalt** . The crust beneath the continents, however, is much more variable in thickness, averaging about 30 kilometers (18.6 miles); under large mountain ranges it can extend to depths of up to 100 kilometers (62.1 miles).

The San Andreas fault is the border between two tectonic plates—the North American Plate and Pacific Plate. Los Angeles is located on the Pacific Plate, and San Francisco is on the North American Plate. In a few million years, the two geographic areas will be right next to each other because the western side of the fault (the Pacific Plate) is moving northward with respect to the rest of the state. The fault is moving at about 2 centimeters (just under an inch) per year.

Continental crust is mostly formed of granite, which is less dense than basalt. This density difference is important in driving the motion of plates, as described below.

Below the crust is the mantle, a dense, hot layer approximately 2,900 kilometers (1,802 miles) thick. At the center of the Earth lies the core, which is composed of an iron–nickel alloy. It is divided into two regions—a liquid outer core and solid inner core. As the Earth rotates, the liquid inner core spins, creating the Earth's magnetic field.

Within the crust and mantle, there also are two important mechanical layers—the lithosphere and asthenosphere. The lithosphere is the outermost of these layers, and comprises the crust and uppermost mantle. The lithosphere is relatively cool, making the rock strong and resistant to deformation. The lithosphere is broken into the moving tectonic (or lithospheric) plates.

Below the lithosphere is a relatively narrow, mobile zone of the mantle called the asthenosphere. The asthenosphere is a weak zone, formed of mostly solid rock (with perhaps a little **magma** mixed in), and flows very slowly, in a manner similar to the ice at a bottom of a glacier. The rigid lithosphere is believed to "float" or move about on the slowly flowing asthenosphere.

Plate Tectonic Theory is Developed

The plate tectonic theory known today evolved in the 1950s, owing to four major scientific developments:

1. Demonstration of the young age of the ocean floor;
2. Confirmation of repeated reversals of the Earth's magnetic field in the geologic past;
3. Emergence of the seafloor-spreading hypothesis and associated recycling of the oceanic crust; and
4. Precise documentation that the Earth's earthquake and volcanic activity was concentrated along subduction zones and mid-ocean ridges.

Youthful Seafloor.

Before the nineteenth century, the depth of the open ocean was a matter of speculation, although most scientists believed it to be flat and featureless. Only in 1855 did the first bathymetric maps reveal the first evidence of underwater mountains in the central Atlantic. In 1947, seismologists found that the sediment layer on the floor of the Atlantic was much thinner than previously thought. Scientists believed that the oceans were over 4 billion years old, and were perplexed by the distinct lack of sediment cover. The answer to this question would prove vital to advancing the theory of plate tectonics.

Magnetic Field Reversals.

In the 1950s, scientists began recognizing magnetic variations in the rocks of the ocean floor. This was not entirely unexpected, since it was known that basalt contained the mineral magnetite, and this mineral was known to locally distort compass readings. In the early part of the twentieth century, geologists recognized that oceanic rocks had normal or reverse polarity (i.e., in normal polarity, the rocks have the same orientation of today's magnetic field). This can be explained by the ability of the magnetite grains to align themselves in the molten basalt with the Earth's

magnetic field. When the rock cools, these grains are "locked" in, recording the magnetic orientation or polarity (normal or reversed) at the time of cooling. As more of the ocean floor was mapped, patterns of alternating stripes of normal and reverse polarity were noted; this became known as magnetic striping.

Seafloor Spreading.

With the discovery of magnetic striping at mid-ocean ridges, scientists began to theorize that mid-ocean ridges mark structurally weak zones where magma from deep within the Earth rises and erupts at the surface. This theory, called seafloor spreading, quickly gained acceptance, but raised an additional question: If new crust is continually being formed at mid-ocean ridges, and the Earth is not increasing in size, what is happening to the old crust? Harry Hess and Robert Dietz postulated that the old crust must be destroyed in the deep canyon-like oceanic trenches, while new crust if formed at the mid-ocean ridges. This theory explained why the Earth is not expanding, there is little sediment on the ocean floor, and oceanic crust is much younger than continental rocks.

Subduction Zones.

The final scientific discovery that cemented the theory of plate tectonics occurred with improvements in seismic detection in the 1950s. Seismologists identified regions of earthquake activity that coincided with Hess's predicted areas of ocean crust generation (mid-ocean ridges) and oceanic lithosphere destruction (subduction zones). Today scientists know that tectonic plates move, because they can measure their motion directly using the global positioning system (GPS).

Plate Tectonic Boundaries

Plate tectonic boundaries are regions where lithospheric plates meet. There are three types of plate tectonic boundaries: divergent, convergent, and transform.

Divergent.

Divergent boundaries occur along spreading ridges where plates are moving apart and new crust is being created by ascending magma from the mantle. An example of a divergent plate boundary is the Mid-Atlantic Ridge. This submerged mountain chain extends from the Arctic to the southern tip of Africa, and is one part of the global ridge system that extends around the Earth. [*] The Mid-Atlantic Ridge spreads at a rate of approximately 2.5 centimeters (1 inch) per year.

Convergent.

Convergent boundaries are regions where lithospheric plates collide. The type of convergence depends on the types of plates involved: namely, (1) oceanic–oceanic convergence; (2) oceanic–continental convergence; (3) continental–continental convergence (see figure).

- *Oceanic–Oceanic.* When two oceanic plates collide, one plate is subducted beneath the other. This occurs as one lithospheric plate becomes older, colder, and denser than the underlying hot, weak asthenosphere. As the lithosphere sinks slowly through the asthenosphere, the uppermost sediments are melted, and the resulting magma reaches the surface to form volcanoes. As a result, subduction zones are marked by an arc of

volcanoes parallel to and about 150 kilometers (93 miles) from the plate margin. An example of oceanic-oceanic collision is the Marianas Trench and the Aleutian Islands in the Pacific Ocean.

- *Oceanic–Continental.* When oceanic and continental plates collide, the oceanic plate is the one that is subducted beneath the continental plate, because the continental crust is lighter and less dense. An example of oceanic-continental collision is seen at the Cascadia Subduction Zone, where the Pacific Plate is being subducted beneath the North American Plate.
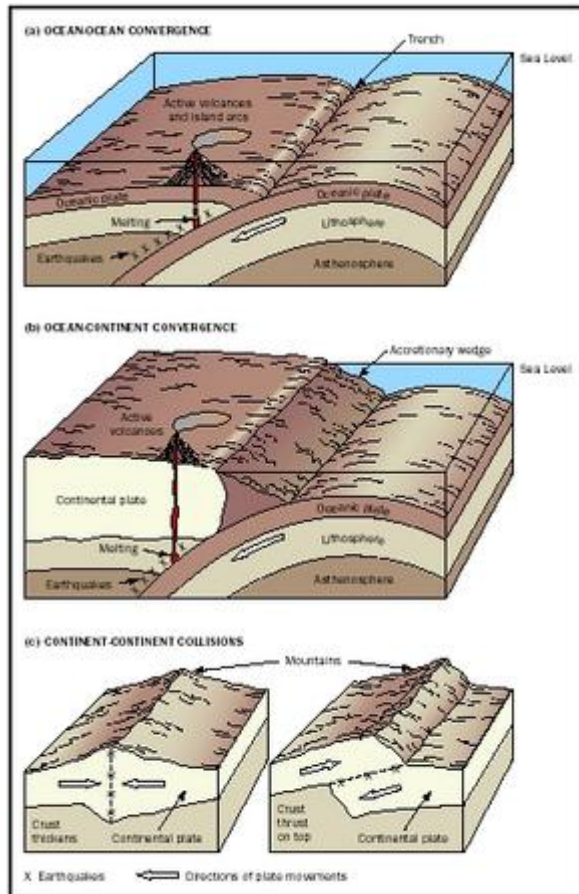
This schematic illustrates the three types of plate convergence.

- *Continental–Continental.* Continental–continental convergence results in spectacular mountain ranges such as the Himalayas, the Alps, and the Appalachians. Because continental crust is buoyant, neither plate will subduct, and a collision zone is the result.

Transform.

Transform boundaries mark regions where plates slide past one another. Transform boundaries are great vertical fractures that extend down through the lithosphere. An example of a transform boundary is the San Andreas Fault in Southern California (see the photograph of the fault on page 202).

Plate Movements

The lithospheric plates do not randomly meander about the Earth's surface, but are driven by internal forces. The mantle is believed to move in circular motions rather like soup boiling on a stovetop, wherein the heated soup rises to the surface, cools, and sinks back to the bottom of the pot, where it is heated and rises again. This cycle is called convective flow, and it is the same process that occurs in the mantle today. However, the heat source within the Earth is radioactive decay of minerals and residual heat from the formation of the Earth.

Until the early 1990s, scientists believed that mantle convection, seafloor spreading, and magma intrusion at mid-ocean ridges (called "ridge push") were the predominant mechanisms that drove plate motion. However, in recent years, the significance of subduction mechanisms over mid-ocean ridge processes has taken precedence. The gravity-controlled sinking of a cold, dense, oceanic slab into a subduction zone (called "slab pull") now is considered the driving mechanism behind plate tectonics.

Although scientists know that forces deep within the Earth drive plate motion, they may never know the exact details, because no mechanism can be directly tested. The fact that lithospheric plates have moved in the past and are still in motion today is beyond dispute, but the exact mechanisms of how and why they move will continue to challenge scientists in the future.

Hot Springs on the Ocean Floor

In many areas of the ocean floor, wherever **magma** nears the seafloor, or where **lava** erupts directly at the seafloor surface, hot springs on the seafloor called hydrothermal vents commonly are found. Vent fields are generally associated with submarine (undersea) volcanoes where lava is erupting or preparing to erupt.

Hydrothermal vents also are found in areas of the ocean floor that are spreading, such as at mid-ocean ridges, where **tectonic plates** are being pulled apart. This movement allows the molten magma to rise from deep inside the Earth, superheating the cold ocean water around it.

The average temperature of deep-ocean water is only 2°C (36°F). The water coming directly from a hydrothermal vent can reach up to 350°C (662°F) and is rich in dissolved chemicals. The hot spring water forms a **plume** above the vent, somewhat like smoke rising from a chimney into the air. Temperature-sensing instruments, towed behind research vessels, can detect these hot-water plumes and aid oceanographers in locating hydrothermal vents on the ocean floor.

Vent Circulation and Chemistry

Seafloor hydrothermal systems influence local ocean chemistry because hydrothermal circulation removes some chemical compounds from sea water, such as magnesium and sulfate, and adds many others. Further, the circulation pattern of sea water within the hydrothermal vent system creates the unique landforms and biological communities found in vent fields.

Sea water enters into the seafloor by seeping down through **fracture** systems. As water percolates through the cracks, it heats as it nears the magma chamber. Subsurface water can heat to 60°C (140°F), warm enough to begin dissolving chemicals and minerals from the surrounding rock. The water becomes rich in chemical compounds of sulfur, iron, and magnesium.

When this vent fluid comes close to the magma chamber, it enters a high temperature reaction zone where temperatures rise to 400°C (752°F). This superheated, high-pressure, mineral-laden vent fluid rises forcefully, searching for fracture systems or other openings to the ocean floor. Hydrothermal
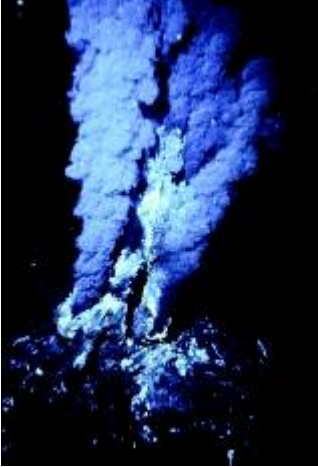
Black smokers form when hot vent water mixes with cold sea water, causing the precipitation of tiny particles of manganese-rich and iron-rich sulfide minerals. White smokers form from slightly cooler vent water and the precipitation of minerals rich in barium and calcium.

vent openings can be as small as several centimeters or as large as a meter or more in diameter. At these openings, the vent fluid becomes focused and expels forcefully into the surrounding cold sea water. At this point, the temperature of the vent fluid varies around 350°C (662°F).

The reaction between the two waters of differing temperature causes some of the dissolved minerals to precipitate out, forming tiny grains in the vent fluid. The particles give the fluid the appearance of a dark cloud. Because a large percentage of the chemical compounds are **sulfides** , such as hydrogen sulfide, the vent fluid is highly toxic, with a **pH** near 4.0, or about that of vinegar.

As the vent fluid rises, it slows and begins to mix with the surrounding ocean water, producing a plume of warm water that starts to drift with the currents. The plume tends to hang together as it cools, aiding in its detection. Many of the chemical and mineral compounds begin to drop from the plume, forming metalliferous sediments and iron and manganese crusts on the nearby ocean floor. Because deposits from hydrothermal vent fluid can contain iron, manganese, copper, zinc, and other minerals, vents have relevance to certain types of ore deposits.

Chimneys, Black Smokers, and Mineral Deposits

"Sulfide chimneys" form at some high-temperature hydrothermal vents. They are made of minerals deposited directly from the vent fluid. These chimneys appear as freestanding hollow tubes with one or more openings to the sea. Chimneys consist of sulfur and iron compounds with a variety of glistening metallic crystal deposits lining the interior. Some sulfide deposits contain quantities of copper, zinc, gold, and silver, leading speculation as to their profitability for mining.

Sulfide chimneys exhibiting no discharge are commonly found, indicating past volcanic activity that has ceased, gone dormant, or shifted underground. These can account for ancient sulfide deposits found, even at sites far from present-day volcanic activity.

There are several types of hydrothermal vents and sulfide chimneys found in a vent field. Some vents appear simply as a hole in the **basalt** rock where a crust of rock has broken away, revealing an open, underground chamber called a lava tube. Others form upright chimneys that expel water, hazy from the high heat and clouded with chemical compounds.

One of the most spectacular of the vents is called a black smoker. It appears as if thick, black smoke is being emitted from the vent. In reality, it is vent fluid filled with tiny mineral grains precipitating when the hot vent fluid comes in contact with the cold sea water.

Chemosynthesis

The chemicals in hydrothermal vent fluid would be toxic to most forms of life familiar to humans; but amazingly, a unique **ecosystem** has evolved to live near hydrothermal vents. The organisms that are supported by the vents rely on **microbes** , similar to colonies of bacteria, which grow in the vent fluid and on the surface of the surrounding rocks and chimneys.

At ocean depths, sunlight is nonexistent and **photosynthesis** is impossible. These organisms are not dependent on sunlight and photosynthesis, but instead rely on chemosynthesis, a process in which certain microbes use chemicals in the vent fluid to produce energy.

Microbes.

In vent communities, chemosynthetic microbes utilize the energy released from hydrogen sulfide **oxidation** to produce food. Essentially, they survive by breaking down the bonds of the hydrogen sulfide molecules, a compound toxic to most life forms. This primary production forms the base for an entire **food chain** of animals dependent upon hydrothermal vents. Chemosynthetic microbes grow on the seafloor, within the fracture systems, and even within other animals at the vents.

Three main types of microbes exist at hydrothermal vents. *Mat-forming microbes* produce a slimy coating that covers all surfaces exposed to the mineral rich vent fluid. The mat consists of many species, from long filamentous microbes to tiny rod-shaped microbes.

Plume microbes grow within the superheated plume of water being expelled by hydrothermal vents. They can survive in a wide range of temperatures, from the deep ocean average of 2°C (36°F) to a high of 110°C (230°F). When expelled from a vent they form a cloud of white flakes giving rise to the name "snow blower vent."

Symbiotic microbes live within the tissues of some animals, producing food for themselves and their host. Symbiotic microbes account for 75 percent of all food producers within the vent community.

Vent Biocommunities

The unique **environment** created by active hydrothermal vents plays host to a web of life found nowhere else on Earth. The ability of microbes to create a food source from otherwise toxic chemicals provides for animal species that biologists are just beginning to understand. Because each species can tolerate certain levels of heat and toxicity, the communities form concentric rings around hydrothermal vents with each species existing in its preferred habitat. Animals also are categorized as vent and non-vent species. Some are directly dependent on the vents for survival while others can migrate between vents and the open ocean.

Tubeworms.

Tubeworms are one of the most common and distinctive animals found at hydrothermal vents. Tubeworms attach themselves to the seafloor and have no mouth or anus. Instead, they have symbiotic microbes living inside them. Tubeworms live in colonies consisting of hundreds to many thousands of individuals, with many other smaller vent species living among them. The colonies form mounds of long, white stalks topped with red, branching filaments.

Other Species.

Numerous species of clams are found near the vents, siphoning the warm water to digest plume microbes and detritus (decaying matter) dropping from above. [*] **Gastropods** such as limpets and snails move about the rocks, rasping the mat-forming microbes with their radulas. Other species of annelid worms and sea cucumbers also feed from the mat-forming microbes. Species of crabs and arthropods scrape away at the microbes as well as prey on the sedentary and slow-moving species. Some fish prefer the vent fields and can tolerate the high heat and toxins, whereas others roam between the vents in search of food. Rarely, a small gray octopus will swim by. Termed *Bentosoctopus,* it is very reclusive and not much is known about the species.
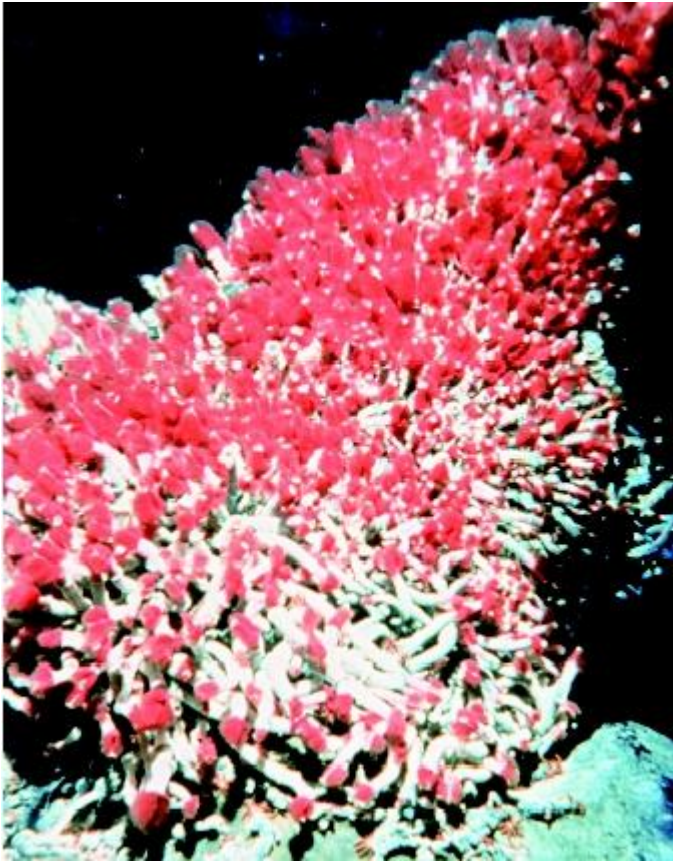
Tubeworms are specially adapted to thrive in vent environments that otherwise would be considered toxic if found in surface waters such as lakes and streams. Here tubeworms feed at the base of a black smoker chimney.

Colonization.

When hydrothermal vents begin to develop, animal communities colonize the vents with a set progression. Plume and mat-forming microbes begin to grow almost immediately. Within a few months, fast-colonizing animals, such as tubeworms, scale worms, snails, and limpets begin to colonize hydrothermal vents. Newly colonized vents typically have few species and few individuals, and if tubeworms are present, they are white and small (generally less than 50 centimeters long, or 19.5 inches). Older, established vents have more species and more animals, and the tubeworms are brown and larger (usually more than 50 centimeters long).

Some sessile (nonmobile) and slow-colonizing animals like sponges, sea fans, and crinoids are mostly found on older lava away from active hydrothermal areas. They are sparsely distributed and colonize new lava flows and vents very slowly, often taking many years. Some species of octopus, crab, and fish are mobile and can move freely from one site to another, and can be found near either old or new lava.

Hydrothermal vents do not remain active forever. Inevitably, the underground magma will cool. Vents begin to cool and go dormant. The nonsessile animals have the ability to migrate to other active vent fields. But those species that are attached or move too slowly will perish as their heat and food source slowly dies out.

Physics and Chemistry of Sea Water

Most people come in contact with the ocean only near its surface, and usually near its edges. In the huge part of the ocean that remains hidden, sea water is salty, cold, dark, and deep. Average salt content in the ocean is 35 grams per kilogram of sea water, composed mostly of six constituents: sodium (Na^+), chloride (Cl^-), sulfate (SO_4^{2-}), magnesium (Mg^{2+}), calcium (Ca^{2+}), and potassium (K^+). [*] These are often referred to as conservative elements, because their ratios to each other remain constant throughout the ocean.

It is important to measure salinity of sea water accurately. The salinity and the temperature determine water density (which drives water movement), and concentrations of many elements can be indirectly determined from salinity.

Salinity

In the early twentieth century, salinity was measured by chemically titrating the sample to measure the chloride (Cl^-) **ion** , then making use of the constant proportions of the major ions to Cl^- to calculate total salinity. In the 1950s and 1960s, it became clear that this procedure and calculation were not satisfactory for the precise measurements needed to distinguish and track different water masses in the oceans.

Drying a sample of sea water and weighing the salt residue is not practical because some of the salts tend to decompose and lose weight before all the water has been removed. The index of refraction of light changes with the salinity of a water sample; a handheld optical instrument called a refractometer can be used to find an inexact measure of salinity.

By 1960, the salinity of sea water was most often measured by a salinometer, which measures the electrical conductivity of the sample compared to that of a standard. Electrical conductivity, the ability of a substance to conduct electricity, increases for water as the amount of dissolved ions increase. Using the measurement of conductivity it is possible to measure salinity with high precision, so salinity can be determined to ±0.0001.

The conductivity can be measured *in situ* (without bringing the water to the surface) at the same time temperature and pressure are measured. These measurements give a profile of salinity and temperature versus depth (pressure) and are commonly collected by an instrument package referred to as a CTD (conductivity, temperature, depth).

The units by which salinity is expressed have changed as the methods of measurement changed. Salinity can be given as grams (g) of salt per kilogram (kg) of sea water (g/kg), or as parts per thousand (ppt). Titration for Cl^- led to the use of chlorinity units, also expressed as ppt. Sea water of 35 ppt salinity has a 19.4 ppt chlorinity.

Salinity is now defined as the ratio of the electrical conductivity of the sample at 15°C and 1 atmosphere of pressure to that of a potassium chloride (KCl) solution containing 32.4356 g KCl in 1 kg of solution. (Atmospheric pressure at sea level is about 14.7 pounds per square inch, termed "1 atmosphere.") The KCl solution is measured at the same temperature and pressure as the sample and gives a conductivity defined to correspond to a practical salinity of 35. Because this is derived from a ratio, it has no units, and is written "salinity is 35."

The above definition was formally established in 1980 by the United Nations Educational, Scientific and Cultural Organization (UNESCO) Joint Panel on Oceanographic Tables and Standards, putting an end to decades of debate about the meaning of salinity. Some authors, uncomfortable with a unitless value, use psu (practical salinity units). In practice, a sample of standard sea water with a salinity of 35.000 is used as a shipboard standard to compare conductivities and calibrate the salinometer.

Temperature and Density

The temperature of the world's ocean is highly variable over the surface of the ocean, ranging from less than 0°C (32°F) near the poles to more than 29°C (84°F) in the tropics. It is heated from the surface downward by sunlight, but at depth most of the ocean is very cold. Seventy-five percent of the water in the ocean falls within the temperature range of –1 to +6°C (30 to 43°F) and the salinity range of 34 to 35.

Variations in total salinity and in temperature cause variations in the density of sea water. Several factors can cause the salinity to deviate from 35. Addition of river water or rainwater decreases salinity; excess evaporation or formation of **pack ice** causes salinity to increase (because ice crystals themselves do not contain salt—the salt is expelled to cracks and pores between the crystals).

Cold sea water is denser than warm sea water. There are several areas at the ocean surface where surface water becomes very cold. In these locations, surface sea water becomes denser than the surrounding water and sinks to begin the formation of slow **thermohaline** currents, which move deep-ocean water.

Density differences among different water masses allow physical oceanographers to calculate movements of water in the ocean. The density of a water sample is a measure of the total mass in a given unit volume. Salinity increases the density because the dissolved salts are contained in the same volume as the water. Water molecules cluster more closely around positive and negative ions in solution in a process called electrostriction, which also serves to increase sea-water density.

Density of water in the ocean, reported as *sigma t* (σ_t), is calculated from temperature, salinity and pressure by using the equation of state for sea water: $\sigma_t = (\sigma - 1) \times 1{,}000$. At 4°C and salinity of 35, the density σ of sea water is 1.02781 gram per cubic centimeter, and $\sigma_t = 27.81$.

At depth, pressure from the overlying ocean water becomes very high (pressure at 4,000 meters is about 400 atmospheres), but water is only slightly compressible, so that there is only a minor pressure effect on density. At a depth of 4,000 meters, water decreases in volume only by 1.8 percent. Although the high pressure at depth has only a slight effect on the water, it has a much greater effect on easily compressible materials (see box on this page).

Density Structure of the Ocean.

The light and heat from the Sun can only directly penetrate a short distance into the ocean. The surface water is warmer and thus less dense than deep water, which gives most of the ocean a stable density arrangement (see the figure on page 82). The temperature and density often are relatively constant in the surface zone or mixed layer (upper 100 to 200 meters), and begin to change more abruptly in deeper water. The mixed layer is only about 2 percent of the total ocean volume, but covers most of its surface.
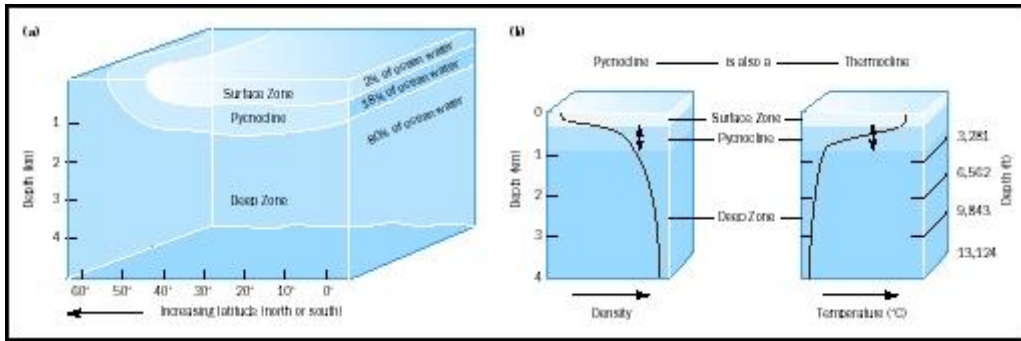
Density stratification (layering) in the ocean is an important feature that influences biological, chemical, and physical processes. As part (a) shows, a shallow surface zone of relatively warm, low-density water overlies a layer called the pycnocline, in which density increases rapidly with depth. Below the pycnocline lies the deep zone of cold, dense water, accounting for about 80 percent of total ocean volume. As part (b) shows, the rapid density increase in the pycnocline is mainly a result of a rapid decrease in temperature (with depth). The pycnocline therefore coincides with the thermocline—a region of rapid temperature change. The pycnocline sometimes may also coincide with the halocline, where salinity increases rapidly with depth; this may occur in shallow water near rivers, for example.

Temperature decreases and density increases more abruptly in deeper water; the same structure exists in lakes. Many swimmers will recall the experience of finding cooler water at greater depths in lakes. The zone separating surface ocean water from deep water is the pycnocline, containing 18 percent of ocean volume. The deep zone lies below the pycnocline, and contains 80 percent of the ocean volume; its temperature and density are much less variable than those found in the pycnocline.

Light and Sound in the Ocean

Most of the ocean lies in complete blackness. Sunlight reaches only depths of about 100 meters (330 feet) in clear open water. This lighted layer is referred to as the photic zone.

The depth of light penetration is decreased by particles in the water, including any algal cells that are growing there. Coastal waters with a high sediment content, or water in which an algal bloom is occurring, have much shallower light penetration than clear open-ocean water. *

Water absorbs different wavelengths of light differently. By a depth of 10 meters (33 feet), mostly blue-green light remains, explaining the bluish color of underwater photos taken in natural light.

Sound.

Sound travels at 1,450 meters (4,750 feet) per second in sea water compared to 334 meters (1,100 feet) per second in air. Sound in water is reflected back when it strikes a solid object. Because the speed of sound in water is well known, this behavior is used to measure distances under the ocean; a signal is sent out and the time required for the return of the reflected sound can give an accurate measure of the distance to the object that reflected it. This technique is used to measure the depth of water from the surface to the seafloor under a ship; the PDR (precision depth recorder) uses a narrow sound beam to give a continuous record of the water depth through which the ship is moving.

Some wavelengths of sound can penetrate the seafloor to some degree, and hence show the layering in sediment. Depth recorders can detect the presence of fish below the surface and record

the movements of the deep scattering layer, swarms of small organisms that move toward and away from the sea surface as the time of day changes.

Sonar (sound navigation and ranging) uses sound to locate and identify targets such as submarines. * Navies have conducted years of research on sound propagation in water. The speed of sound in water increases as temperature, salinity, and depth increase. Differences in these properties in layers of ocean water cause the sound to refract, or bend, as it travels through the ocean. Refraction can easily make the sound appear to come from a different direction than its real source location, so an accurate understanding of sound physics has been vital for naval operations.

A zone of minimum sound velocity exists at a depth of roughly 1,000 meters (about 0.6 mile) called the sofar (sound fixing and ranging) channel. Sound signals that originate in the sofar channel tend to stay in the channel rather than escaping. The sound may travel enormous distances in this channel; explosions set off in the channel in Australia have been heard in Bermuda.

A project called ATOC (Acoustic Thermometry of Ocean Climate) was being tested as of 2002. Its goal is to measure global climate change by observing changes in the speed of sound in the sofar channel that would indicate changes in ocean temperatures. Research to determine the effects of this project on marine mammals also is underway.

Freezing of Sea Water

The density of ocean water is determined by its *salinity* (or salt content) and temperature. The saltier and/or colder the water is, the denser it is. Salt water is most dense at its freezing point, unlike fresh water, which is most dense at about 3.9°C (39.0°F). Oceans are highly stratified: deep ocean water is heavy, and the lighter water is on top. This often is not the case in the atmosphere: warm air near the ground surface is lighter than the overlying air it displaces, resulting in unstable conditions and thunderstorms.

The stratified, stable nature of oceans is important because otherwise there would be no **sea ice** , nor would there be warm tropical seas. Tropical oceans are temperature-stratified: a **thermocline** separates the warm, light water on top from the frigid deep waters. Polar oceans, on the other hand, are salinity-stratified: the salinity is slightly lower on top, especially in the Arctic and near the **estuaries** of large rivers such as the MacKenzie in Canada and the Ob in Russia. This fresh-water outflow will stay on top of the Arctic Ocean and mix only very slowly, because it is lighter. Instead of a thermocline, there is a halocline, a layer in which the salinity changes rapidly with depth. [*]

Salty water freezes below 0°C (32°F): this is why salt is used to melt the snow or ice on a road pavement. The saltier the **brine** , the lower its freezing point. This is also why salt traditionally was added to the water–ice mixture used to make ice cream.

Ocean water with a typical salinity of 35 parts per thousand freezes only at −1.8°C (28.9°F). So if there were no halocline in the polar oceans, then the cooled top ocean layer, being denser, would sink into the deep ocean, in same way as thunderstorm clouds rise in the atmosphere, and the entire ocean column would have to cool to −1.8°C before its surface could freeze.

Formation of Sea Ice

As sea water freezes, salt is excluded, because salt has a different crystalline structure: it forms cubic crystals (with four sides) whereas ice is hexagonal, or six-sided. (A close look at tiny snowflakes will reveal their hexagonal form.) So pockets of brine form within the ice; they refuse to freeze, because of the high salinity. The brine then slowly leaches out of the bottom of the forming ice and drips into the ocean below. Thus sea ice, when melted, is considerably fresher than the original sea water from which it formed. This brine rejection process creates dense water below the ice surface. This layer of dense water may sink to the bottom of the ocean. Furthermore, this

Ice floes and other forms of sea ice are less salty than the sea water from which they formed, owing to a process known as brine rejection. Yet sea ice still is too salty to be melted for human consumption. Only icebergs, which are derived from glaciers, are composed of fresh-water ice. process of "deep-water formation" is an exception to the normally stratified nature of oceans.

The salt enrichment is most effective when it occurs in a thin layer on a calm sea. The faster sea ice is frozen, the less salt that can escape. For instance, when frozen at an air temperature of −40°C (−40°F), the salinity of the ice is about 10 percent. But when frozen at an air temperature of −6°C (21°F), the salinity of the ice is only about 4 percent. Such ice is fresh enough to use as drinking water; in fact, in spring, polar bears often drink the water in melting ponds on ice.

When sea ice melts in the summer, the meltwater forms a relatively fresh surface layer that lies above the saltier ocean water, maintaining the halocline, and allowing easier freezing the next winter.

Sound Transmission in the Ocean

The speed of sound depends on the medium through which sound waves propagate. The speed of sound differs in air and water, with sound waves traveling faster in water. For example, in air at a temperature of 18°C (64°F), the speed of sound is approximately 341 meters (1,120 feet) per second. In contrast, in salt water at approximately the same temperature, the speed of sound is approximately 1,524 meters (5,000 feet) per second.

The crew of *Le Suroit* prepare to launch deep-sea sonar in 1985 during a search for the wreck of the *Titanic.* Sonar (sound navigation and ranging) uses sound to locate and identify targets, whether shipwrecks, submarines, or schools of fish.

The state properties of water (temperature and pressure) and the degree of salinity also affect the speed of sound. The propagation of sound waves in sea water can be directly affected by suspensions of particulate matter that can scatter, absorb, or reflect the waves. Laboratory experiments demonstrate that distilled water—water from which salts and other suspended particles have been removed—provides a medium in which the speed of sound exceeds the speed of sound in ocean water. The difference in the speed of transmission is significant—speed in distilled water may be 20 to 30 times that of speeds found in ocean water.

Because frequency and wavelength are inversely proportional characteristics of sound waves, low-frequency signals produce long sound wavelengths. These long-wavelength signals encounter fewer suspended particles as they pass through the medium and thus are not as subject to scattering, absorption, or reflection. As a result, low-frequency signals are able to travel farther

without significant loss of signal strength. Naval communication systems utilize low-frequency, long-wavelength signals to enhance communications with submerged submarines.

Physical Differences Produce a "Sound Channel"

Within the ocean, the speed of sound varies with changes in depth that accompany normal changes in temperature and pressure. Specific combinations of temperature, pressure, and salinity may act to create shadow zones, or reflective layers, that are resistant to the propagation of sound waves. A specific set of conditions, however, also act to create a channel through which sound waves propagate at minimal speed but with minimal loss of strength. Similar to the transmission of light through fiber-optic cables, the refraction (bending) of sound waves by layers of water with varying temperature, pressure, and/or salinity allow the formation of a well-defined sound channel.

Although the oceans are not uniform bodies of water—there are currents of water with dramatic variations of temperature (such as the Gulf Stream) and salinity—the speed of sound in the deeper regions of the oceans is influenced more by high pressure. Conversely, at shallower depths, temperature plays the most dominant role in governing the speed of sound.

The greater the temperature of the water, the faster sound travels. Surface temperature variation can be significant with seasonal variations in the amount of sunlight (insolation) that can produce changes in near-surface temperatures that, in turn, affect the speed of sound in water near the ocean surface. When the near-surface layer is well mixed by currents and surface action, a resulting isothermal layer allows uniform propagation speeds for sound waves. Such isothermal layers are common in mid-latitude regions.

The SOFAR Channel

A temperature gradient exists when the temperature of the water decreases with increasing depth. The resulting **thermocline** shows a characteristic decrease in the speed of sound with decreasing temperature. However, at a depth of approximately 750 meters (2,460 feet), the variations in temperature become so slight that the water becomes essentially isothermal (of uniform temperature). From that point, the speed of sound is regulated more by changes in pressure that accompany the increasing depth.

Because sound wave transmission speed is directly proportional to pressure, the speed of sound increases as the pressure increases with depth. Accordingly, at the interface of the thermocline and the isothermal depths, there exists a region of minimal speed of sound. This interface creates a sound "pipeline," or "deep sound channel," within the oceans that allows the transmission of low-frequency sound over thousands of kilometers. This sound fixing and ranging (SOFAR) channel was discovered in 1943 by an American team led by Maurice Ewing and J. L. Worzel, and independently by Soviet physicist Leonid Brekhovskikh. Ewing and Worzel demonstrated that the SOFAR channel was capable of transmitting the low-frequency, long-wavelength sound waves produced by an explosion near the Bahama Islands to receivers stationed near the coast of Africa.

SOFAR channel depths also are a function of the depth and thickness (extent) of the thermocline. For example, SOFAR channels run nearer the surface in colder, polar seas.

Temperature and pressure affect water density, and the refraction of sound waves occurs at the interface of mediums (or layers within a medium) of differing density. Because of refraction, sound waves traveling through the SOFAR channel are deflected toward a region of lower velocity. Accordingly, waves traveling upward toward the surface, where the speed of sound

increases with increasing temperature, deflect downward. Waves traveling toward deeper water, where the speed of sound increases with increasing pressure, are deflected upward.

Sound waves can be trapped effectively in the narrow SOFAR channel. Traveling at minimum velocity, the sound waves lose little energy, allowing the waves to propagate over distances in excess of 25,000 kilometers (15,500 miles).

Prior to the widespread use of GPS (global positioning system) equipment, the SOFAR channel also was used for navigation and the location of marine craft. Some scientists hypothesize that certain species of whales utilize the SOFAR channel to communicate mating calls over long distances.

Based on the known relationship of temperature changes to changes in the speed of sound, the Acoustic Thermometry of Ocean Climate (ATOC) project is attempting to provide data crucial to measurement of changes in global temperature. By measuring differences in the speed of sound transmitted over long distances (such as across the Pacific basin), data accumulated over a long timeframe should average out variations in temperature and salinity, enabling ATOC scientists to calculate changes in ocean temperature that may provide evidence related to questions regarding global warming.

Light Transmission in the Ocean

Visible radiation, or light, from the Sun is important to the world's ocean systems for several reasons. It provides the energy necessary for ocean currents and wind-driven waves. Conversion of some of that energy into heat helps form the thin layer of warm water near the ocean's surface that supports the majority of marine life. Most significantly, the transmission of light in sea water is essential to the productivity of the oceans.

Visible wavelengths of light are captured by chlorophyll-bearing marine plants, which then make their own food through the process of **photosynthesis** . The organic molecules created by this process are an important energy source for many small organisms that are the base of the entire marine **food chain** . All life in the oceans is ultimately dependent upon the light and the process of photosynthesis that it initiates. Similarly, light transmission is a key factor in the **ecology** of lakes and streams, which are discussed elsewhere in this encyclopedia.

Reflection, Refraction, and Color

The uppermost, sunlit layer of the ocean where 70 percent of the entire amount of photosynthesis in the world takes place is called the euphotic zone. It generally extends to a depth of 100 meters (330 feet). Below this is the disphotic zone, between 100 and 1,000 meters (330 and 3,300 feet) deep, which is dimly lit. Some animals are able to survive here, but no plants. Although the amount of light is measurable at this range of depths, there is not enough available for photosynthesis to take place. The layer of the ocean where no light at all penetrates—over 90 percent of the entire ocean area on Earth—is called the aphotic zone, where depths are more than 1,000 meters (3,300 feet).

Light Penetration.

A certain amount of incoming light is reflected away when it reaches the ocean surface, depending upon the state of the water itself. If it is calm and smooth, less light will be reflected.

If it is turbulent, with many waves, more light will be reflected. The light that penetrates the surface is **refracted** due to the fact that light travels faster in air than in water. Once it is within the water, light may be scattered or absorbed by solid particles. Most of the visible light spectrum is absorbed within 10 meters (33 feet) of the water's surface, and almost none penetrates below 150 meters (490 feet) of water depth, even when the water is very clear.

Greater abundances of solid particles in the water will decrease the depth of light penetration. Therefore, water near the seashore that is more turbid (cloudy) due to particles will show a decrease in light transmission, even in shallow water. This is due to large numbers of particles brought in by river systems, and biological production by **microorganisms** , as well as waves, tides, and other water movement picking up debris on the ocean floor.

Light Spectrum.

Water selectively scatters and absorbs certain wavelengths of visible light. The long wavelengths of the light spectrum—red, yellow, and orange—can penetrate to approximately 15, 30, and 50 meters (49, 98, and 164 feet), respectively, while the short wavelengths of the light spectrum—

violet, blue and green—can penetrate further, to the lower limits of the euphotic zone. Blue penetrates the deepest, which is why deep, clear ocean water and some tropical water appear to be blue most of the time. Moreover, clearer waters have fewer particles to affect the transmission of light, and scattering by the water itself controls color. Water in shallow coastal areas tends to contain a greater amount of particles that scatter or absorb light wavelengths differently, which is why sea water close to shore may appear more green or brown in color.

Moorings and Platforms

Moorings and platforms are structures that allow water scientists to position instruments, collect samples, and take long-term measurements in waterbodies. A mooring typically consists of a flexible cable that is tethered to the ocean floor by a weight or anchor and suspended from the sea surface by a buoy. A platform has a flat workspace for instruments and people, and can be free-floating, moored to the seafloor, or rigidly fixed in place.

Water scientists, including physical and chemical oceanographers, marine biologists and geologists, **meteorologists** , and lake scientists called limnologists collect many types of data from instruments installed on moorings

Instruments that measure physical and chemical aspects of the oceans and atmosphere can mounted on or suspended from a buoy. Together with shipboard and satellite data, the global network of measuring sites provides further understanding of the ocean-atmosphere system and its influence on global climate and biological productivity.

and platforms. Oceanographers measure physical and chemical properties such as temperature, salinity, current velocity and turbidity—the amount of suspended sediment and other particles in the water—using equipment suspended beneath the sea surface on mooring cables. Marine biologists collect water and specimen samples, conduct dives, and make observations of marine flora and fauna from platforms. They also use oceanographic information to study the

distribution of marine **ecosystems** . Marine geologists collect seafloor samples and sub-seafloor cores that hold clues to the Earth's geologic history and past climate, as well as clues to the location of **petroleum reservoirs** , from fixed platforms.

Equipment and Instrumentation

Moorings and platforms often accommodate numerous pieces of measuring equipment. This economical approach allows scientists to collect several types of data at a site, to compare data from the site, and to observe data patterns that affect a number of natural systems. Changes in current direction or temperature within the water column, for example, may correlate to changes in biological productivity or weather patterns.

A typical oceanographic mooring, like one deployed in the northwest Atlantic Ocean by the Global Ocean Ecoystems Dynamics (GLOBEC) program, holds a large array of instrumentation: seven current meters, seven temperature gauges, three optical turbidity scanners, four salinity/conductivity/pressure meters, and one Acoustic Doppler Current Profiler (ADCP) that records surface ocean current patterns around the mooring. The GLOBEC mooring buoy houses instruments that record wind speed and direction, air temperature, relative humidity, and amount of sunlight reaching the buoy at the sea surface.

Deployment.

Tethered moorings and platforms are deployed from ships and left at their sites to record data for extended periods of time. The GLOBEC moorings were deployed for 5 years and serviced every 6 months. In the past, ocean scientists had to retrieve data from recorders at such "stand alone" moorings during return visits to the sites. Modern moorings, however, now include transmitters that beam data to scientists' computers via satellite. They also carry global positioning system (GPS) instruments that measure their precise location over time, and transmitters that relay information about mechanical difficulties at moorings. Some systems can even receive signals to adjust their instruments. These technological advances allow continuous data monitoring at land stations, improve data quality, and reduce the number of expensive visits to mooring sites.

An international initiative called the Global Ocean Observing System (GOOS) is a global scientific collaboration that seeks to collect data from moorings and platforms all over the planet, and to compile those data into a global database that would be available to scientists for comparison and analysis.

Specialized Design.

Tethered moorings and platforms are designed to reliably collect, record, and transmit accurate data while submerged in corrosive saltwater and exposed to extreme weather conditions for long periods of time. Their buoys have heavy-duty steel or aluminum flotation collars, resilient solar panels, batteries for backup power, safety devices such as beacons and radio transmitters, and insulated, reinforced wells that protect sensitive scientific instruments.

Mooring cables, while flexible, are extremely strong. Electronic cables that carry data from the submerged instruments to the sea surface are encased in an insulating sheath that surrounds the tether. Anchor weights, which are designed to hold moorings in place indefinitely, can be mechanically or electronically signaled to release when it is time to retrieve the mooring.

Platforms

Platforms provide flat workspaces to accommodate scientists and large equipment at the study site and serve as field laboratories for water scientists. In addition to collecting the types of data provided by unmanned moorings,

The data collected at moorings and platforms give scientists the information they need to pursue some of the most pressing scientific and environmental questions of the twenty-first century. Here, oceanographers with the National Oceanic and Atmospheric Administration prepare floats for a mooring deployment from the deck of a research vessel in Antarctica's Ross Sea.

staffed platforms allow scientists to collect biological specimens, sea water, and sediment samples, to observe marine organisms, to conduct diving expeditions, and to deploy to manned and unmanned submersible vehicles. Instruments installed on drifting platforms measure oceanographic, atmospheric, and biological data along the path of an ocean current.

Geological studies that seek to collect deep sediment cores, or to drill into crystalline rock, require platforms that are stable in shifting seas. Some platforms float around a central "moon pool" that isolates the drilling equipment from platform motion. Other drilling platforms stand above the sea surface on legs that are rooted in the seafloor.

Most drilling platforms are too expensive for publicly funded scientific studies, but they are common equipment in the petroleum industry. Oil platforms over major petroleum reservoirs may house hundreds of staff members who live and work on the rig for weeks or months at a time. The petroleum industry also maintains smaller, unmanned platforms that house oil and gas pumps once an offshore field is established.

Oceanography from Space

The use of space satellite data for ocea n observations allows marine scientists to view biological, chemical, and physical interactions within the oceans on regional and global scales. Satellite studies have revolutionized our ideas of how the ocean works. Satellite sensors measure a myriad of different phenomena including: sea surface temperature, surface wind, ocean color and productivity, ocean height, tides, and currents.

Altimeter Data

Several different in struments are used to collect oceanographic measurements. For example, the TOPEX/Poseidon satellite measures the height of the sea surface using an **altimeter** . The altimeter flies aboard the satellite at approximately 1,335 kilometers (830 miles) above Earth and emits a radar pulse that reflects off the sea surface. Because the speed of the pulse and location of the satellite are known, scientists can calculate the height of the sea surface.

In addition, the strength of the radar signal depends on the size of the surface ripples, which are in turn related to the wind speed, allowing the wind speed to be calculated. Because currents are detectable as slopes in the sea surface, the world's ocean currents can also be identified and monitored.

Altimeter data is also used for identifying the **topography** of the seafloor. For example, when there is a topographic high, such as a mountain on the seafloor, then there is a related topographic high or "mountain" in the sea level. This seafloor feature can be a subsurface **seamount** , or it may be a local increase in density in the Earth's crust.

Radiometer Data

A radiometer can be used to collect surface temperature data and ocean color data. A radiometer measures the amount of the Sun's visible light and infrared radiation reflected off the ocean. Examples of radiometers include the Coastal Zone Color Scanner (CZCS) and SeaWiFS (Sea-viewing Wide Field-of-view Sensor).

The temperature of the sea surface can be calculated using the infrared portion of the data. Because currents and water masses vary considerably in

This view from the space shuttle *Columbia* shows silt running into the sea from the Mahakam River in Borneo, Indonesia, and the delta that has formed as a result. The delta is the roughly triangular-shaped landmass extending from where the river branches into many distributaries, out to the coastal area. The feathery areas seaward of the delta are very fine-grained sediments that are being transported away by coastal ocean currents.

temperature, this data is particularly useful in observing currents and circulation processes.

Eddies are one feature in particular that can be identified using an infrared radiometer. These are generated by large-scale currents, such as the Gulf Stream. Eddies can affect the distribution of marine life and can last for many years before dissipating. Locating such eddies and studying their dynamics can help researchers track pollution such as oil spills and determine where marine life may be located.

Ocean Color.

Ocean color can be determined by measuring the portions of the visible spectrum reflected from the ocean surface. It can indicate a number of things to an oceanographer, such as amount of **plankton** and amount of vegetation. The color of the ocean changes slightly, from a bright blue to a dark blue or black when plankton float freely or concentrate in areas. These concentrations are called blooms. These colors can indicate to scientists the productivity of the oceans and potential for greater amounts of wildlife since plankton are the basis of the marine food web and without plankton all marine life would suffer.

Satellite data have become accurate and dependable enough that they is now integrated with other forms of marine data collection. In addition, satellite data provide a large-scale view of ocean dynamics that otherwise would be unavailable. What has emerged is exciting new information about vast areas of previously unstudied open water.