Chapter 3

The Physical Setting

Earth is a prolate ellipsoid, an ellipse of rotation, with an equatorial radius of $R_e = 6,378.1349$ km (West, 1982) which is slightly greater than the polar radius of $R_p = 6,356.7497$ km. The small equatorial bulge is due to Earth's rotation.

Distances on Earth are measured in many different units, the most common are degrees of latitude or longitude, meters, miles, and nautical miles. *Latitude* is the angle between the local vertical and the equatorial plane. A meridian is the intersection at Earth's surface of a plane perpendicular to the equatorial plane and passing through Earth's axis of rotation. *Longitude* is the angle between the standard meridian and any other meridian, where the standard meridian is that which passes through a point at the Royal Observatory at Greenwich, England. Thus longitude is measured east or west of Greenwich.

A degree of latitude is not the same length as a degree of longitude except at the equator. Latitude is measured along great circles with radius R, where R is the mean radius of Earth. Longitude is measured along circles with radius $R \cos \varphi$, where φ is latitude. Thus 1° latitude = 111 km; and 1° longitude = 111 cos φ km. For careful work, remember that Earth is not a sphere, and latitude varies slightly with distance from the equator. The values listed here are close enough for our discussions of the oceans.

Because distance in degrees of longitude is not constant, oceanographers measure distance on maps using degrees of latitude.

Nautical miles and meters are connected historically to the size of Earth. Gabriel Mouton, who was vicar of St. Paul's Church in Lyons, France, proposed in 1670 a decimal system of measurement based on the length of an arc that is one minute of a great circle of Earth. This eventually became the nautical mile. Mouton's decimal system eventually became the metric system based on a different unit of length, the meter, which was originally intended to be one tenmillionth the distance from the Equator to the pole along the Paris meridian. Although the tie between nautical miles, meters, and Earth's radius was soon abandoned because it was not practical, the approximations are still useful. For example, the polar circumference of Earth is approximately $2\pi R_e = 40,075$ km. Therefore one ten-thousandth of a quadrant is 1.0019 m. Similarly, a nautical



Figure 3.1 The Atlantic Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the ETOPO 30' data set. The 200 m contour outlines continental shelves.

mile should be $2\pi R_e/(360 \times 60) = 1.855$ km, which is very close to the official definition of the *international nautical mile*: 1 nm $\equiv 1.852$ km.

3.1 Oceans and Seas

There are only three oceans by international definition: the Atlantic, Pacific, and Indian Oceans (International Hydrographic Bureau, 1953). The seas, which are part of the ocean, are defined in several ways, and we will consider two.

The Atlantic Ocean extends northward from Antarctica and includes all of the Arctic Sea, the European Mediterranean, and the American Mediterranean more commonly known as the Caribbean sea (Figure 3.1). The boundary between the Atlantic and Indian Oceans is the meridian of Cape Agulhas (20°E). The boundary between the Atlantic and Pacific Oceans is the line forming the shortest distance from Cape Horn to the South Shetland Islands. In the north,



Figure 3.2 The Pacific Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the ETOPO 30' data set. The 200 m contour outlines continental shelves.

the Arctic Sea is part of the Atlantic Ocean, and the Bering Strait is the boundary between the Atlantic and Pacific.

The Pacific Ocean extends northward from Antarctica to the Bering Strait (Figure 3.2). The boundary between the Pacific and Indian Oceans follows the line from the Malay Peninsula through Sumatra, Java, Timor, Australia at Cape Londonderry, and Tasmania. From Tasmania to Antarctica it is the meridian of South East Cape on Tasmania 147°E.

The Indian Ocean extends from Antarctica to the continent of Asia including the Red Sea and Persian Gulf (Figure 3.3). Some authors use the name Southern Ocean to describe the ocean surrounding Antarctica.

Mediterranean Seas are mostly surrounded by land. By this definition, the Arctic and Caribbean Seas are both Mediterranean Seas, the Arctic Mediterranean and the Caribbean Mediterranean.



Figure 3.3 The Indian Ocean viewed with an Eckert VI equal-area projection. Depths, in meters, are from the ETOPO 30' data set. The 200 m contour outlines continental shelves.

Marginal Seas are defined by only an indentation in the coast. The Arabian Sea and South China Sea are marginal seas.

3.2 Dimensions of the Oceans

The oceans and adjacent seas cover 70.8% of the surface of the Earth, which amounts to $361,254,000 \text{ km}^2$. The areas of the oceans vary considerably (Table 3.1), and the Pacific is the largest.

Oceanic dimensions range from around 1500 km for the minimum width of the Atlantic to more than 13,000 km for the north-south extent of the Atlantic and the width of the Pacific. Typical depths are only 3–4 km. So horizontal dimensions of ocean basins are 1,000 times greater than the vertical dimension. A scale model of the Pacific, the size of an 8.5×11 in sheet of paper, would have dimensions similar to the paper: a width of 10,000 km scales to 10 in, and a depth of 3 km scales to 0.003 in, the typical thickness of a piece of paper.

Table 3.1 Surface Area of the Oceans †

Lable off Surface fire	c of the occans
Pacific Ocean	$181.34\times10^6~{\rm km^2}$
Indian Ocean	$74.12\times10^6~{\rm km^2}$
Atlantic Ocean	$106.57\times 10^6~{\rm km^2}$

[†] From Menard and Smith (1966)



Figure 3.4 Cross-section of the South Atlantic along 25° S showing the continental shelf offshore of South America, a seamount near 35° W, the mid-Atlantic Ridge near 14° W, the Walvis Ridge near 6° E, and the narrow continental shelf off South Africa. **Upper** Vertical exaggeration of 180:1. **Lower** Vertical exaggeration of 30:1. If shown with true aspect ratio, the plot would be the thickness of the line at the sea surface in the lower plot.

Because the oceans are so thin, cross-sectional plots of ocean basins must have a greatly exaggerated vertical scale to be useful. Typical plots have a vertical scale that is 200 times the horizontal scale (Figure 3.4). This exaggeration distorts our view of the ocean. The edges of the ocean basins, the continental slopes, are not steep cliffs as shown in the figure at 41°W and 12°E. Rather, they are gentle slopes dropping down 1 meter for every 20 meters in the horizontal.

The small ratio of depth to width of ocean basins has dynamical implications. Vertical velocities must be much smaller than horizontal velocities. Even over distances of a few hundred kilometers, the vertical velocity must be on order 1% of the horizontal velocity. We will use this information later to simplify the equations of motion.

At first glance, the relatively small value of vertical velocities seems to have little influence on dynamics until we begin to think about turbulence. Three dimensional turbulence is very different than two-dimensional turbulence. In two dimensions, vortex lines must always be vertical, and there can be little vortex stretching. In three dimensions, vortex stretching plays a fundamental role in turbulence.

3.3 Bathymetric Features

Earth's crust is divided into two types: regions of thin dense crust with thickness of about 10 km, the oceanic crust; and regions of thick light crust with thickness of about 40 km, the continental crust. The deep, lighter continental crust floats



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Figure 3.5 Left Histogram of elevations of land and depth of the sea floor as percentage of area of the Earth, in 50 m intervals showing the clear distinction between continents and sea floor. Right Cumulative frequency curve of height, the hypsographic curve. The curves are calculated from the ETOPO 30' data set.

higher on the denser mantle than does the oceanic crust, and the mean height of the crust relative to sea level has two distinct values: continents have a mean elevation of 1114 m; oceans have a mean depth of -3432 m (Figure 3.5).

The volume of the water in the oceans exceeds the volume of the ocean basins, and some water spills over on to the low lying areas of the continents. These shallow seas are the continental shelves. Some, such as the South China Sea, are more than 1100 km wide. Most are relatively shallow, with typical depths of 50–100 m. A few of the more important shelves are: the East China Sea, the Bering Sea, the North Sea, the Grand Banks, the Patagonian Shelf, the Arafura Sea and Gulf of Carpentaria, and the Siberian Shelf. The shallow seas help dissipate tides, they are often areas of high biological productivity, and they are usually included in the exclusive economic zone of adjacent countries.

The crust is broken into large plates that move relative to each other. New crust is created at the mid-ocean ridges, and old crust is lost at trenches. The relative motion of crust, due to plate tectonics, produces the distinctive features of the sea floor sketched in Figure 3.6, include mid-ocean ridges, trenches, island arcs, basins, and sea mounts.

The names of the subsea features have been defined by the International Hydrographic Bureau (1953), and the following definitions are taken from Dietrich et al. (1980).



Figure 3.6 Schematic section through the ocean showing principal features of the sea floor. Note that the slope of the sea floor is greatly exaggerated in the figure.

Basins are depressions of the sea floor more or less equidimensional in form and of variable extent.

Canyon are relatively narrow, deep depressions with steep slopes, the bottoms of which grade continuously downward.

Continental (or island) shelfs are zones adjacent to a continent (or around an island) and extending from the low-water line to the depth at which there is usually a marked increase of slope to greater depth. (Figure 3.7)

Continental (or island) slopes are the declivities seaward from the shelf edge into greater depth.

Plains are flat, gently sloping or nearly level regions of the sea floor, e.g. an abyssal plain.

Ridges are long, narrow elevations of the sea floor with steep sides and irregular topography.

Seamounts are isolated or comparatively isolated elevations rising 1000 m or more from the sea floor and of limited extent across the summit (Figure 3.8).

Sills are the low parts of the ridges separating ocean basins from one another or from the adjacent sea floor.

Trenches are long, narrow, and deep depressions of the sea floor, with relatively steep sides (Figure 3.9).

Subsea features have important influences on the ocean circulation. Ridges separate deep waters of the oceans into distinct basins separated by sills. Water deeper than a sill cannot move from one basin to another. Tens of thousands of isolated peaks, seamounts, are scattered throughout the ocean basins. They interrupt ocean currents, and produce turbulence leading to vertical mixing of water in the ocean.



Figure 3.7 An example of a continental shelf, the shelf offshore of Monterey California showing the Monterey and other canyons. Canyons are are common on shelfs, often extending across the shelf and down the continental slope to deep water. Figure copyright Monterey Bay Aquarium Research Institute (MBARI).

3.4 Measuring the Depth of the Ocean

The depth of the ocean is usually measured two ways: 1) using acoustic echosounders on ships, or 2) using data from satellite altimeters.

Echo Sounders Most maps of the ocean are based on measurements made by echo sounders. The instrument transmits a burst of 10–30 kHz sound and listens for the echo from the sea floor. The time interval between transmission of the pulse and reception of the echo, when multiplied by the velocity of sound, gives twice the depth of the ocean (Figure 3.10).

The first transatlantic echo soundings were made by the U.S. Navy Destroyer Stewart in 1922. This was quickly followed by the first systematic survey of a ocean basin, made by the German research and survey ship Meteor during its expedition to the South Atlantic from 1925 to 1927. Since then, oceanographic and naval ships have operated echo sounders almost continuously while at sea. Millions of miles of ship-track data recorded on paper have been digitized to produce data bases used to make maps. The tracks are not well distributed. Tracks tend to be far apart in the southern hemisphere, even near Australia (Figure



Figure 3.8 An example of a seamount, the Wilde Guyot. A guyot is a seamount with a flat top created by wave action when the seamount extended above sea level. As the seamount is carried by plate motion, it gradually sinks deeper below sea level. The depth was contoured from echo sounder data collected along the ship track (thin straight lines) supplemented with side-scan sonar data. Depths are in units of 100 m.

- 3.11) and closer together in well mapped areas such as the North Atlantic. Depths measured by echo sounders are useful, but they do have errors:
 - 1. Sound speed varies by $\pm 4\%$ in different regions of the ocean. Tables of the mean sound speed are used to correct depth measurements to an accuracy of around $\pm 1\%$. See §3.6 for more on sound in the ocean.
 - 2. Echoes may come from shallower depths off to the side of the ship rather from directly below the ship. This can introduce small errors in some hilly regions.
 - 3. Ship positions were poorly known before the introduction of satellite navigation techniques in the 1960s. Ship positions could be in error by tens of kilometers, especially in cloudy regions where accurate celestial fixes could not be obtained.
 - 4. Schools of marine zooplankton or fish were sometimes mistaken for shallow water, leading to false seamounts and shoals on some bathymetric charts. This error is reduced by remapping questionable features.



Figure 3.9 An example of a trench, the Aleutian Trench; an island arc, the Aleutian Islands; and a continental shelf, the Bering Sea. The island arc is composed of volcanos produced when oceanic crust carried deep into a trench melts and rises to the surface. **Top:** Map of the Aleutian region of the North Pacific. **Bottom:** Cross-section through the region.

5. Some oceanic areas as large as 500 km on a side have never been mapped by echo sounders (Figure 3.11). This creates significant gaps in knowledge of the oceanic depths.

Satellite Altimetry Gaps in our knowledge of ocean depths between ship tracks have now been filled by satellite-altimeter data. Altimeters profile the shape of the sea surface, and it's shape is very similar to the shape of the sea floor. To see this, we must first consider how gravity influences sea level.

The Relationship Between Sea Level and the Ocean's Depth Excess mass at the seafloor, for example the mass of a seamount, increases local gravity because the mass of the seamount is larger than the mass of water it displaces, rocks



Figure 3.10 Left: Echo sounders measure depth of the ocean by transmitting pulses of sound and observing the time required to receive the echo from the bottom. **Right:** The time is recorded by a spark burning a mark on a slowly moving roll of paper. (From Dietrich, et al. 1980)



Figure 3.11 Locations of echo-sounder data used for mapping the ocean near Australia. Note the large areas where depths have not been measured from ships. (From Sandwell)

being more than three times denser than water. The excess mass increases local gravity, which attracts water toward the seamount. This changes the shape of the sea surface (Figure 3.12).

Let's make the concept more exact. To a very good approximation, the sea surface is a particular *level surface* called the *geoid* (see box). By definition a level surface is everywhere perpendicular to gravity. In particular, it must be perpendicular to the local vertical determined by a plumb line, which is a line from which a weight is suspended. Thus the plumb line is perpendicular to the local level surface, and it is used to determine the orientation of the level surface, especially by surveyors on land.

The excess mass of the seamount attracts the plumb line's weight, causing the plumb line to point a little toward the seamount instead of toward Earth's center of mass. Because the sea surface must be perpendicular to gravity, it must have a slight bulge above a seamount as shown in the figure. If there were no bulge, the sea surface would not be perpendicular to gravity. Typical seamounts produce a bulge that is 1–20 m high over distances of 100–200 kilometers. Of course, this bulge is too small to be seen from a ship, but it is easily measured by an altimeter. Oceanic trenches have a deficit of mass, and they produce a depression of the sea surface.

The correspondence between the shape of the sea surface and the depth of the water is not exact. It depends on the strength of the seafloor and the age of the seafloor feature. If a seamount floats on the seafloor like ice on water, the gravitational signal is much weaker than it would be if the seamount rested on the seafloor like ice resting on a table top. As a result, the relationship between gravity and bathymetry varies from region to region.

Depths measured by acoustic echo sounders are used to determine the regional relationships. Hence, altimetery is used to interpolate between acoustic echo sounder measurements (Smith and Sandwell, 1994). Using this technique, the ocean's depth can be calculated with an accuracy of ± 100 m.

Satellite-altimeter systems Now lets see how altimeters can measure the shape of the sea surface. Satellite altimeter systems include a radar to measure the height of the satellite above the sea surface and a tracking system to determine the height of the satellite in geocentric coordinates. The system measures the height of the sea surface relative to the center of mass of the Earth (Figure 3.13). This gives the shape of the sea surface.

Many altimetric satellites have flown in space. All have had sufficient accuracy to observe the marine geoid and the influence of bathymetric features on the geoid. Typical accuracy varied from a few meters for GEOSAT to ± 0.05 m for Topex/Poseidon. The most useful satellites include Seasat (1978), GEOSAT (1985–1988), ERS–1 (1991–1996), ERS–2 (1995–), and Topex/Poseidon (1992–). Seasat, ERS–1, and ERS–2 also carried instruments to measure winds, waves, and other processes. GEOSAT and Topex/Poseidon are primarily altimetric satellites.

Satellite Altimeter Maps of the Bathymetry Seasat, GEOSAT, ERS-1, and ERS-2 were operated in orbits designed to map the marine geoid. Their orbits had ground tracks spaced 3–10 km apart, which is sufficient to map the geoid. The first measurements, which were made by GEOSAT, were classified by the

The Geoid

The level surface corresponding to the surface of an ocean at rest is a special surface, the *geoid*. To a first approximation, the geoid is an ellipsoid that corresponds to the surface of a rotating, homogeneous fluid in solid-body rotation, which means that the fluid has no internal flow. To a second approximation, the geoid differs from the ellipsoid because of local variations in gravity. The deviations are called *geoid undulations*. The maximum amplitude of the undulations is roughly ± 60 m. To a third approximation, the geoid deviates from the sea surface because the ocean is not at rest. The deviation of sea level from the geoid is defined to be the *topography*. The definition is identical to the definition for land topography, for example the heights given on a topographic map.

The ocean's topography is caused by tides and ocean surface currents, and we will return to their influence in chapters 10 and 18. The maximum amplitude of the topography is roughly ± 1 m, so it is small compared to the geoid undulations.

Geoid undulations are caused by local variations in gravity, which are due to the uneven distribution of mass at the sea floor. Seamounts have an excess of mass due to their density and they produce an upward bulge in the geoid (see below). Trenches have a deficiency of mass, and they produce a downward deflection of the geoid. Thus the geoid is closely related to bathymetry; and maps of the oceanic geoid have a remarkable resemblance to the bathymetry.



Figure 3.12 Seamounts are more dense than sea water, and they increase local gravity causing a plumb line at the sea surface (arrows) to be deflected toward the seamount. Because the surface of an ocean at rest must be perpendicular to gravity, the sea surface and the local geoid must have a slight bulge as shown. Such bulges are easily measured by satellite altimeters. As a result, satellite altimeter data can be used to map the sea floor. Note, the bulge at the sea surface is greatly exaggerated, a two-kilometer high seamount would produce a bulge of approximately 10 m.

US Navy, and they were not released to scientists outside the Navy. By 1996 however, the geoid had been mapped by the Europeans and the Navy released all the GEOSAT data. By combining data from all altimetric satellites, the small errors due to ocean currents and tides have been reduced, and maps of the geoid with ± 3 km spatial resolution have been produced.



Figure 3.13 A satellite altimeter measures the height of the satellite above the sea surface. When this is subtracted from the height r of the satellite's orbit, the difference is sea level relative to the center of the Earth. The shape of the surface is due to variations in gravity, which produce the geoid undulations, and to ocean currents which produce the oceanic topography, the departure of the sea surface from the geoid. The reference ellipsoid is the best smooth approximation to the geoid. (From Stewart, 1985).

3.5 Bathymetric Charts and Data Sets

Most available echo-sounder data have been digitized and plotted to make bathymetric charts. Data have been further processed and edited to produce digital data sets which are widely distributed in CD-ROM format. These data have been supplemented with data from altimetric satellites to produce maps of the sea floor with spatial resolution approaching 3 km.

The British Oceanographic Data Centre publishes the General Bathymetric Chart of the Oceans (GEBCO) Digital Atlas on behalf of the Intergovernmental Oceanographic Commission of UNESCO and the International Hydrographic Organization. The atlas consists primarily of the location of bathymetric contours, coastlines, and tracklines from the GEBCO 5th Edition published at a scale of 1:10 million. The original contours were drawn by hand based on digitized echo-sounder data plotted on base maps.

The U.S. National Geophysical Data Center publishes the ETOPO-5 CD-ROM containing values of digital oceanic depths from echo sounders and land heights from surveys interpolated to a 5-minute (5-nautical mile) grid. Much of the data were originally compiled by the U.S. Defense Mapping Agency, the U.S. Navy Oceanographic Office, and the U.S. National Ocean Service. Although the map has values on a 5-minute grid, data used to make the map are much more sparse, especially in the southern ocean, where distances between ship tracks can exceed 500 km in some regions. The same data set and CD-ROM is contains values smoothed and interpolated to a 30-minute grid.

Sandwell and Smith of the Scripps Institution of Oceanography distribute a digital bathymetric atlas of the oceans based on measurements of the height of the sea surface made from GEOSAT and ERS-1 altimeters and echo-sounder data. This map has a spatial resolution of 3–4 km and a vertical accuracy of



Figure 3.14 The bathymetry of the ocean with 3 km resolution produced from satellite altimeter observations of the shape of the sea surface (From Smith and Sandwell).

 ± 100 m (Smith and Sandwell, 1997). The US National Geophysical Data Center combined the Sandwell and Smith data with land elevations to produce a global map with 2 minute spatial resolution. These maps shows much more detail than the ETOPO-5 map because the satellite data fill in the regions between ship tracks (Figure 3.14).

National governments publish coastal and harbor maps. In the USA, the NOAA National Ocean Service publishes nautical charts useful for navigation of ships in harbors and offshore waters.

3.6 Sound in the Ocean

Sound provides the only convenient means for transmitting information over great distances in the ocean, and it is the only signal that can be used for the remotely sensing of the ocean below a depth of a few tens of meters. Sound is used to measure the properties of the sea floor, the depth of the ocean, temperature, and currents. Whales and other ocean animals use sound to navigate, communicate over great distances, and to find food.

Sound Speed The sound speed in the ocean varies with temperature, salinity, and pressure (MacKenzie, 1981; Munk *et al.* 1995: 33):

$$C = 1448.96 + 4.591 T - 0.05304 T^{2} + 0.0002374 T^{3} + 0.0160 Z$$
(3.1)
+ (1.340 - 0.01025 T)(S - 35) + 1.675 × 10⁻⁷ Z - 7.139 × 10⁻¹³ T Z^{3}

where C is speed in m/s, T is temperature in Celsius, S is salinity in practical salinity units (see Chapter 6 for a definition of salinity), and Z is depth in meters. The equation has an accuracy of about 0.1 m/s (Dushaw, *et al.* 1993). Other



Figure 3.15 Processes producing the sound channel in the ocean. Left: Temperature T and salinity S measured as a function of depth during the R.V. *Hakuho Maru* cruise KH-87-1, station JT, on 28 January 1987 at Lat $33^{\circ}52.90'$ N, Long $141^{\circ}55.80'$ E in the North Pacific. Center: Variations in sound speed due to variations in temperature, salinity, and depth. Right: Sound speed as a function of depth showing the velocity minimum near 1 km depth which defines the sound channel in the ocean. (Data from JPOTS Editorial Panel, 1991).

sound-speed equations have been widely used, especially an equation proposed by Wilson (1960) which has been widely used by the U.S. Navy.

For typical oceanic conditions, C varies within a small range, typically within 1450 m/s to 1550 m/s (Fig. 3.13). Using (3.1), we can calculate the sensitivity of C to changes of temperature, depth, and salinity typical of the ocean. The approximate values are: 40 m/s per 10°C rise of temperature, 16 m/s per 1000 m increase in depth, and 1.5 m/s per 1 psu increase in salinity. Thus the primary causes of variability of sound speed is temperature and depth (pressure). Variations of salinity are too small to have much influence.

If we plot sound speed as a function of depth, we find that the speed usually has a minimum at a depth around 1000 m (Figure 3.16). The depth of minimum speed is called the *sound channel*. It occurs in all oceans, and it usually reaches the surface at very high latitudes.

The sound channel has great practical importance. Refraction allows sound produced at this depth to propagate to great distances. Sound rays that begin to travel out of the channel are refracted back toward the center of the channel. Rays propagating upward at small angles to the horizontal are bent downward, and rays propagating downward at small angles to the horizontal are bent upward (Figure 3.16). Typical depths of the channel vary from 10 m to 1200 m depending on geographical area.



Figure 3.16 Ray paths of sound in the ocean for a source near the axis of the sound channel. (From Munk *et al.* 1995)

Absorption of Sound Absorption of sound per unit distance depends on the intensity I of the sound:

$$dI = -kI_0 \, dx \tag{3.2}$$

where I_0 is the intensity before absorption and k is an absorption coefficient which depends on frequency of the sound. The equation has the solution:

$$I = I_0 \exp(-kx) \tag{3.3}$$

Typical values of k (in decibels dB per kilometer) are: 0.08 dB/km at 1000 Hz; and 50 dB/km at 100,000 Hz. Decibels are calculated from: $dB = 10 \log(I/I_0)$. where I_0 is the original acoustic power, I is the acoustic power after absorption.

For example, at a range of 1 km a 1000 Hz signal is attenuated by only 1.8%: $I = 0.982I_0$. At a range of 1 km a 100,000 Hz signal is reduced to $I = 10^{-5}I_0$. In particular the 30,000 Hz signal used by typical echo sounders to map the ocean's depth are little attenuated going from the surface to the bottom and back.

Very low frequency sounds in the sound channel, those with frequencies below 500 Hz have been detected at distances of megameters. In 1960 15-Hz sounds from explosive charges detonated in the sound channel off Perth Australia were heard in the sound channel near Bermuda, nearly halfway around the world. Later experiment showed that 57-Hz signals transmitted in the sound channel near Heard Island (75°E, 53°S) could be heard at Bernuda in the Atlantic and at Monterey, California in the Pacific (Munk *et al.* 1994).

Use of Sound Because low frequency sound can be heard at great distances, the US Navy, in the 1950s placed arrays of microphones on the seafloor in deep and shallow water and connected them to shore stations. The Sound Surveillance System SOSUS, although designed to track submarines, has found many other uses. It has been used to listen to and track whales up to 1,700 km away, and to find the location of subsea volcanic eruptions.

3.7 Important Concepts

- 1. If the oceans were scaled down to a width of 8 inches they would have depths about the same as the thickness of a piece of paper. As a result, the velocity field in the ocean is nearly 2-dimensional. Vertical velocities are much smaller than horizontal velocities.
- 2. There are only three official oceans.
- 3. The volume of ocean water exceeds the capacity of the ocean basins, and the oceans overflow onto the continents creating continental shelves.
- 4. The depths of the ocean are mapped by echo sounders which measure the time required for a sound pulse to travel from the surface to the bottom and back. Depths measured by ship-based echo sounders have been used to produce maps of the sea floor. The maps have poor spatial resolution in some regions because the regions were seldom visited by ships and ship tracks are far apart.
- 5. The depths of the ocean are also measured by satellite altimeter systems which profile the shape of the sea surface. The local shape of the surface is influenced by changes in gravity due to subsea features. Recent maps based on satellite altimeter measurements of the shape of the sea surface combined with ship data have depth accuracy of ± 100 m and spatial resolutions of ± 3 km.
- 6. Typical sound speed in the ocean is 1480 m/s. Speed depends primarily on temperature, less on pressure, and very little on salinity. The variability of sound speed as a function of pressure and temperature produces a horizontal sound channel in the ocean. Sound in the channel can travel great distances; and low-frequency sounds below 500 Hz can travel halfway around the world provided the path is not interrupted by land.