I am indebted to colleagues at the University of Rhode Island who looked at different parts of this 2nd edition while it was under development: Dave Herbert, Randy Watts and Mark Wimbush; and to several outside reviewers who looked at the entire manuscript: Frank Bub, Kevin Leaman, Allan Robinson and Robert Smith. I am particularly grateful to Ray Najjar and his students at Pennsylvania State University where an earlier version of this manuscript was used in class. He and his students provided extensive suggestions. This 2nd edition is better for all of this review. Its limitations and faults are mine.

— John A. Knauss

CHAPTER ONE

Overview

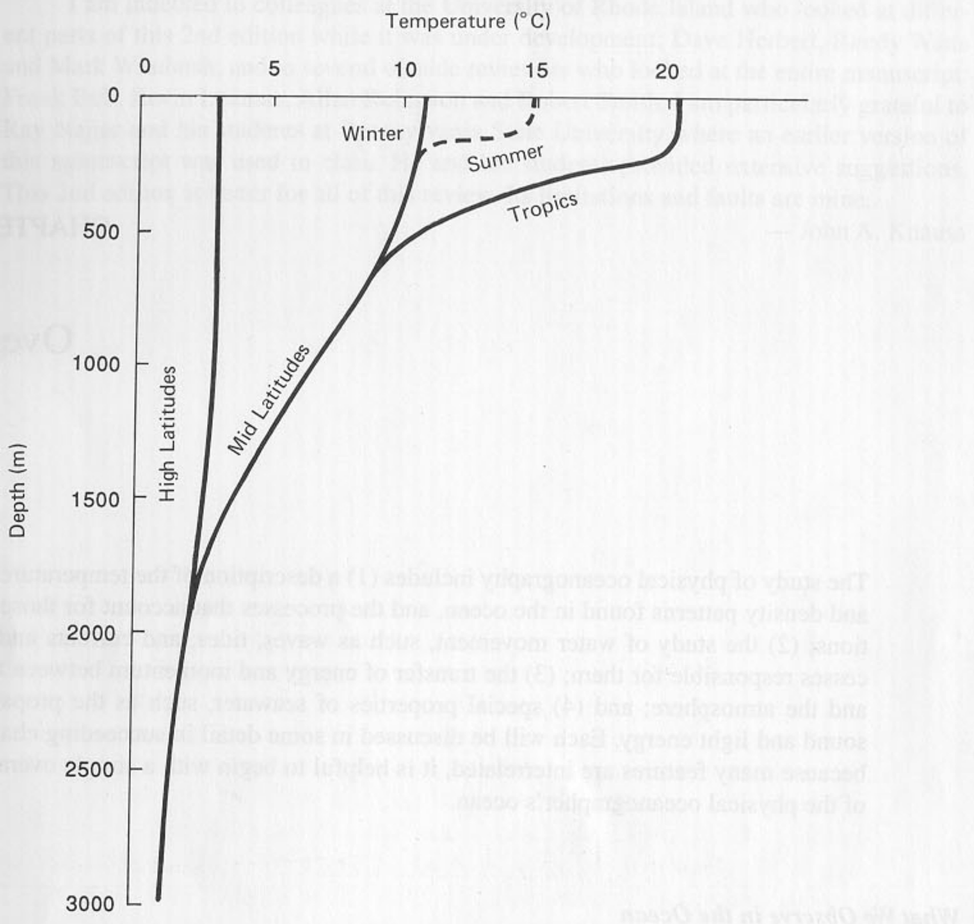
The study of physical oceanography includes (1) a description of the temperature, salinity, and density patterns found in the ocean, and the processes that account for those distributions; (2) the study of water movement, such as waves, tides, and currents and the processes responsible for them; (3) the transfer of energy and momentum between the ocean and the atmosphere; and (4) special properties of seawater, such as the propagation of sound and light energy. Each will be discussed in some detail in succeeding chapters, but because many features are interrelated, it is helpful to begin with a simple overall picture of the physical oceanographer's ocean.

What We Observe in the Ocean

Much of physical oceanography concerns the distribution of temperature, salinity, and density throughout the ocean.

Temperature. With few exceptions, the temperature of the ocean decreases with depth. Generally, the decrease is more rapid near the surface than at depth. A typical temperature-versus-depth profile has a surface layer tens of meters thick, generally referred to as the *mixed layer,* because surface winds usually play an important role in keeping the water well mixed and maintaining a nearly isothermal condition. Below the mixed layer is a region of rapidly changing temperature referred to as the *thermocline.*

Below the thermocline the temperature decreases slowly with depth, becoming nearly isothermal again (Figure 1.1). The water in the lower half of all oceans is uniformly cold. Some 75% of the water in the ocean has a temperature between 0 and 4°C (Figure 1.2, Table 1.1). The discovery that the deep water of the tropical ocean was cold was made late in the eighteenth century, and the obvious conclusion was soon drawn: The deep water must originate at polar latitudes. A typical longitudinal cross section shows a thin layer of



igure1.1 Typical temperature profiles in the open ocean. Below a relatively shallow surface layer, he ocean is uniformly cold.

varm water confined to middle and low latitudes, with cold water at depth and at high latudes (Figure 1.3).

An important reason for the warm water of the ocean being confined to shallow epths is that unlike the atmosphere, the ocean is an effective absorber of the sun's radiaon. More than 50% of the heat from the sun makes it through the atmosphere to the surace of the earth. More than 50% of that which reaches the sea surface is absorbed within he first meter of the ocean, and in even the clearest of midocean waters less than 1% of he sun's energy penetrates to 100 m. In a typical coastal region, where the water is discolred with high biological productivity or with sediment stirred up from the bottom, the 9% absorption level can be reached in less than 10 m (Figure 1.4).

The ocean, like the land, is heated in summer and cooled in winter, but because the un's heat is absorbed so near the surface, seasonal changes in the ocean temperature ructure are mostly confined to a relatively shallow surface layer. The warm surface water

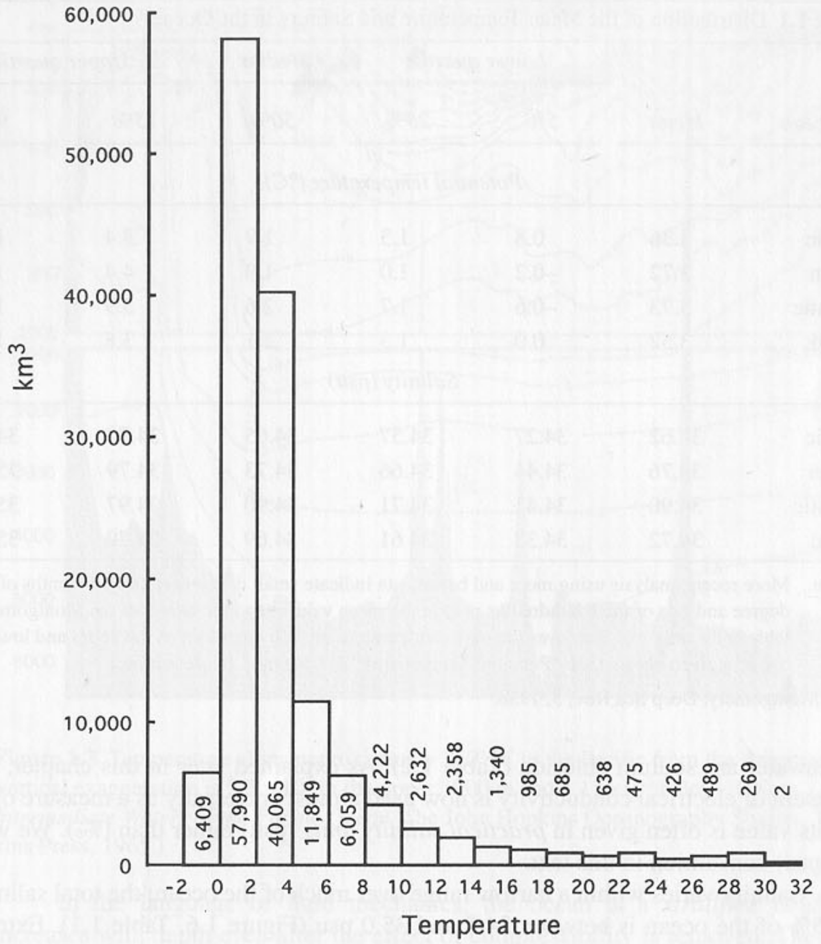


Figure 1.2 The distribution of temperature in the world's oceans. The histogram represents the number of cubic kilometers of seawater in each 2°C temperature range. Approximately 75% of the ocean has a temperature between 0°C and 4°C. (After Montgomery, *Deep-Sea Res.*,*5*, 1958.)

of summer is less dense than the underlying cooler water and is not easily mixed downward. Seasonal changes in the temperature of the ocean are seldom observed below 200 m (Figure 1.5). This is in marked contrast to the summer atmosphere, in which the sun warms the earth, which in turn warms the surface air, which expands and becomes lighter as it warms and is convected upward to form the summer cumulus clouds often several kilometers above the surface.

Salinity. The total amount of dissolved material in seawater is its salinity. It is given in parts per thousand, grams of salt per kilogram of seawater, and in much of the oceanographic literature one finds the symbol (‰) representing parts per thousand. All of the chemical elements found on earth can be found in seawater, but 87% of the dissolved salts

Table 1.1 Distribution of the Mean Temperature and Salinity in the Ocean[a]

Ocean	Mean	*Lower quartile*		*Median*	*Upper quartile*	
		5%	*25%*	*50%*	*75%*	*95%*
		Potential temperature (°C)				
Pacific	3.36	0.8	1.3	1.9	3.4	11.1
Indian	3.72	–0.2	1.0	1.9	4.4	12.7
Atlantic	3.73	–0.6	1.7	2.6	3.9	13.7
World	3.52	0.0	1.3	2.1	3.8	12.6
		Salinity (psu)				
Pacific	34.62	34.27	34.57	34.65	34.70	34.79
Indian	34.76	34.44	34.66	34.73	34.79	35.19
Atlantic	34.90	34.41	34.71	34.90	34.97	35.73
World	34.72	34.33	34.61	34.69	34.79	35.10

a. More recent analysis using more and better data indicate small differences (several tenths of a degree and two or three hundredths psu) in the mean values reported here, but the Montgomery table is the only one that gives the mean temperature of such quantities as the upper and lower 5% of a given ocean. (See "Potential Temperature" in Chapter 2 for definition.)

After Montgomery, Deep Sea Res., *5*, 1958.

in seawater are sodium chloride (Table 1.2). As explained later in this chapter, the measurement of electrical conductivity is now used almost universally as a measure of salinity, and its value is often given in *practical salinity units* (psu) rather than (‰). We will adopt the (psu) convention in this text.

Salinity varies within a narrow range over much of the ocean; the total salinity range of 75% of the ocean is between 34.5 and 35.0 psu (Figure 1.6, Table 1.1). Extremes are found in confined evaporation basins such as the Red Sea (high) and off major fresh-water river outflows (low). For many purposes, we can simply assume that the ocean is of constant salinity. However, our understanding of the details of ocean processes is often predicated on small salinity differences. For example, the salinity of water in the deep Pacific (below 2500 m) changes from 34.70 psu in the South Pacific to 34.68 psu at 40° N. Oceanographers agree that this small change can best be explained by assuming that this water moves slowly northward and is being diluted by less saline water above it, which is being mixed downward.

***Density*.** The density of seawater is determined by its temperature, salinity, and depth (in effect, the hydrostatic pressure to which it is subjected). Density increases with decreasing temperature and increasing salinity and pressure. Hydrostatic pressure increases about 1 atmosphere (1 bar, 10^5 Pascal, [Pa]) every 10 m in depth. The compressed water at 4000 m with the same temperature and salinity has a density of about 1046 kg/m^3 compared to a value of 1028 kg/m^3 at the surface.

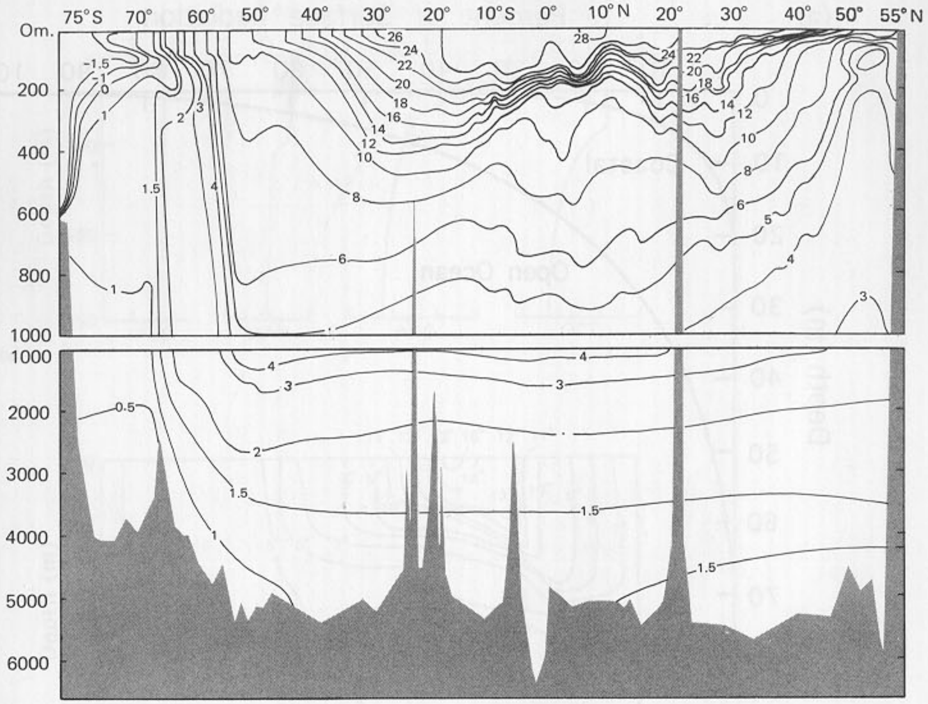


Figure 1.3 Temperature along approximately 160° W in the Pacific from the Antarctic to Alaska. Vertical exaggeration is 5.5×10^3 in the upper 1000 m and 1.11×10^3 below 1000 m. (After Reid, *Intermediate Waters of the Pacific Ocean,* The John Hopkins Oceanography Studies, The John Hopkins Press, 1965.)

In the language of fluid mechanics, the ocean is a *stratified fluid.* Its density increases with depth even after the effect of compressibility is removed. The difference in density between the surface and the deep ocean (after the effect of compressibility has been removed) is only a few parts in a thousand (e.g., the difference between 1025 and 1028 kg/m^3). This may appear to be a relatively small difference, but this small amount of stratification exerts a powerful influence on the processes one observes in the ocean.

Scale. The oceans cover about 70% of the earth and have an average depth slightly less than 4000 m. Tables 1.3 and 1.4 and Figure 1.7 give some useful figures. We tend to draw cross sections of oceanic properties with considerable vertical exaggeration (see, for example, Figure 1.3), but it is important to remember that a typical ocean width is measured in thousands of kilometers and depths are in thousands of meters. The true ratio of ocean area to depth more closely resembles the ratio of area to thickness of this page; thus vertical exaggerations of 1000:1 or more, as in Figure 1.3, are the rule rather than the exception. Similar exaggeration occurs in thinking about bottom slopes. The apparently precipitous plunge down the continental slope from the shelf to the oceanic abyss has a typical slope of 4% (for an example, see Figure 7.4).

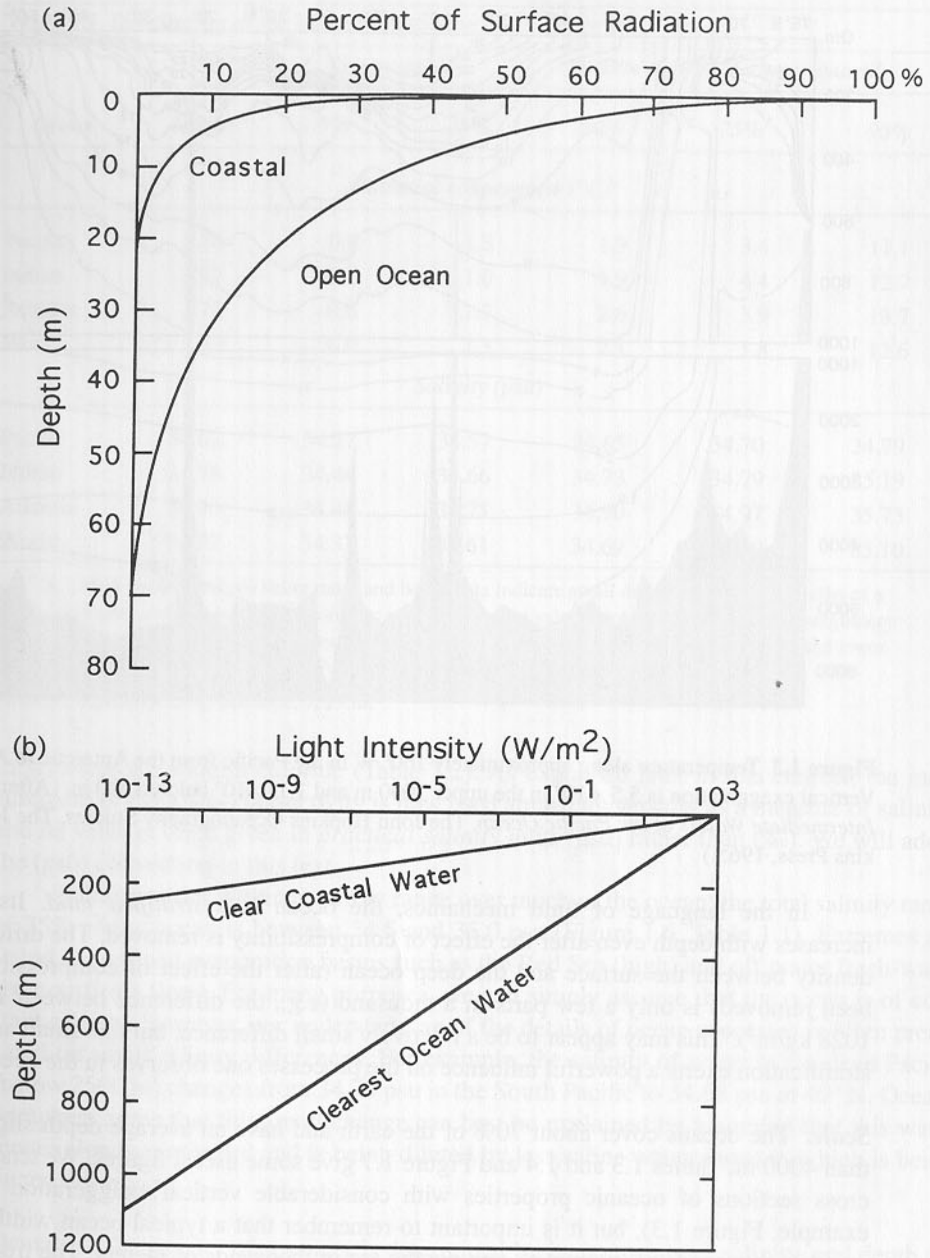


Figure 1.4 (a) The percentage of light energy that reaches a given depth. (b) Similar information plotted in terms of light intensity.

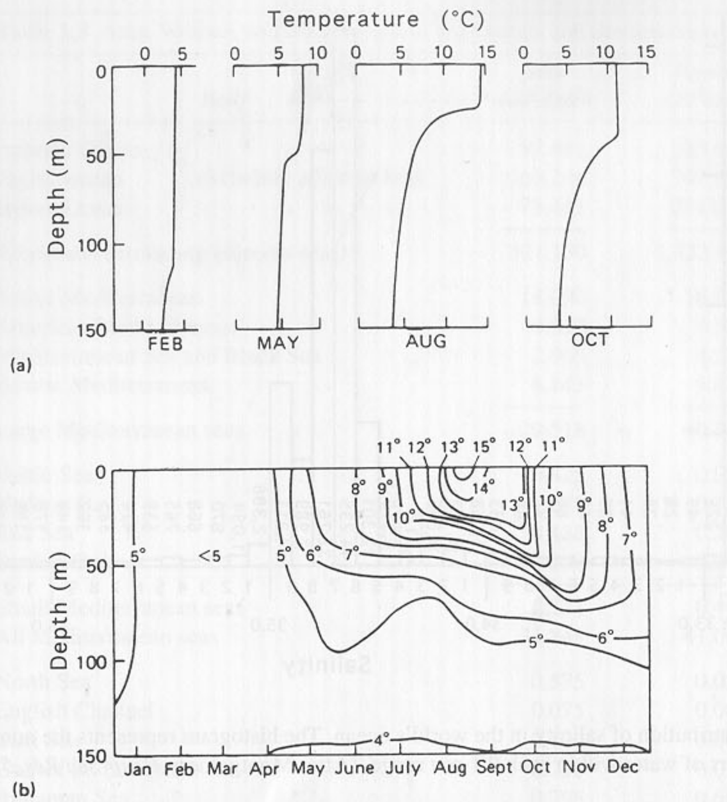


Figure 1.5 (a) Typical seasonal changes in temperature structure in the top 150 m at a Northern Hemisphere midlatitude site. (b) Comparable data in a time-versus-depth plot. The summer build-up and winter destruction of the thermocline can be seen clearly.

Ocean Circulation

It is convenient, at least as one begins a study of physical oceanography, to divide the circulation of the ocean into *wind-driven currents* and *thermohaline circulation.* The former, as the name implies, are generated primarily by wind. The primary energy source for the latter is the sun. However, since the sun is the primary energy source for the winds of the atmosphere, one can also argue that it is the sun which ultimately provides the energy for both types of ocean circulation.

Wind-Driven Currents. The mean wind fields over the both the North and South Atlantic and Pacific are characterized by westerlies (winds out of the west) at midlatitudes and the easterly trade winds at low latitudes. The frictional drag of these winds imparts a spin to the surface waters, clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere, creating the great current gyres that one observes in these oceans (Figure 1.8). The geometry of the ocean basins and the details of the mean wind fields in

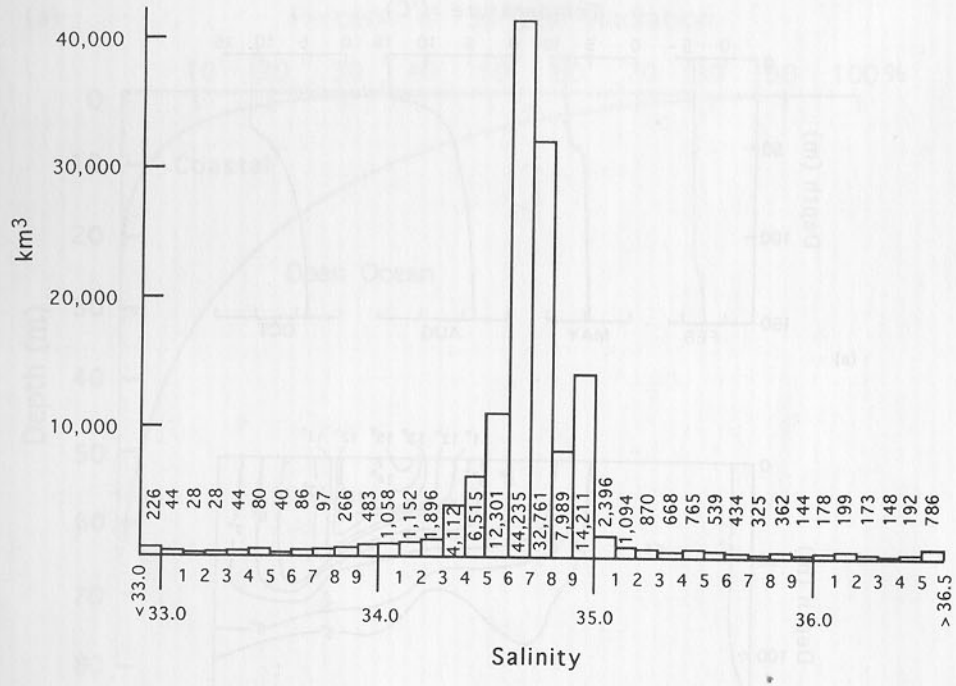


Figure 1.6 Distribution of salinity in the world's ocean. The histogram represents the number of cubic kilometers of water within each 0.1 psu range. (After Montgomery, *Deep-Sea Res.*, *5*, 1958.)

Table 1.2 Major Constituents in Seawater

Constituent	*g/kg in Seawater*[a]	*Percentage by Weight*	
Chloride	19.35	55.07	
Sodium	10.76	30.62	
Sulfate	2.71	7.72	
Magnesium	1.29	3.68	99.36
Calcium	0.41	1.17	
Potassium	0.39	1.10	
Bicarbonate	0.14	0.40	
Bromide	0.067	0.19	
Strontium	0.008	0.02	
Boron	0.004	0.01	
Fluoride	0.001	0.01	
Total		99.99	

a. Salinity = 35‰

After Riley and Skirrow, *Chemical Oceanography,* Academic Press, 1975.

Table 1.3 Area, Volume, and Mean Depth of the Oceans and Semienclosed Seas

Body	*Area* *(10^6 km^2)*	*Volume* *($10^6 km^3$)*	*Mean depth* *(m)*
Atlantic Ocean (excluding adjacent seas)	82.441	323.613	3926
Pacific ocean (excluding adjacent seas)	165.246	707.555	4282
Indian Ocean (excluding adjacent seas)	73.443	291.030	3963
All oceans (excluding adjacent seas)	321.130	1,322.198	4117
Arctic Mediterranean	14.090	16,980	1205
American Mediterranean	4.319	9.573	2216
Mediterranean Sea and Black Sea	2.966	4.238	1429
Asiatic Mediterranean	8.143	9.873	1212
Large Mediterranean seas	29.518	40.664	1378
Baltic Sea	0.422	0.023	55
Hudson Bay	1.232	0.158	128
Red Sea	0.438	0.215	491
Persian Gulf	0.239	0.006	25
Small Mediterranean seas	2.331	0.402	172
All Mediterranean seas	31.849	41.066	1289
North Sea	0.575	0.054	94
English Channel	0.075	0.004	54
Irish Sea	0.103	0.006	60
Gulf of St. Lawrence	0.238	0.030	127
Andaman Sea	0.798	0.694	870
Bering Sea	2.268	3.259	1437
Okhotsk Sea	1.528	1.279	838
Japan Sea	1.008	1.361	1350
East China Sea	1.249	0.235	188
Gulf of California	0.162	0.132	813
Bass Strait	0.075	0.005	70
Marginal seas	8.079	7.059	874
All adjacent seas	39.928	48.125	1205
Atlantic Ocean (including adjacent seas)	106.463	354.679	3332
Pacific Ocean (including adjacent seas)	179.679	723.699	4028
Indian Ocean (including adjacent seas)	74.917	291.945	3897
All oceans (including adjacent seas)	361.059	1370.323	3795

After Sverdrup, Johnson, and Fleming, *The Oceans,* Prentice Hall, 1942.

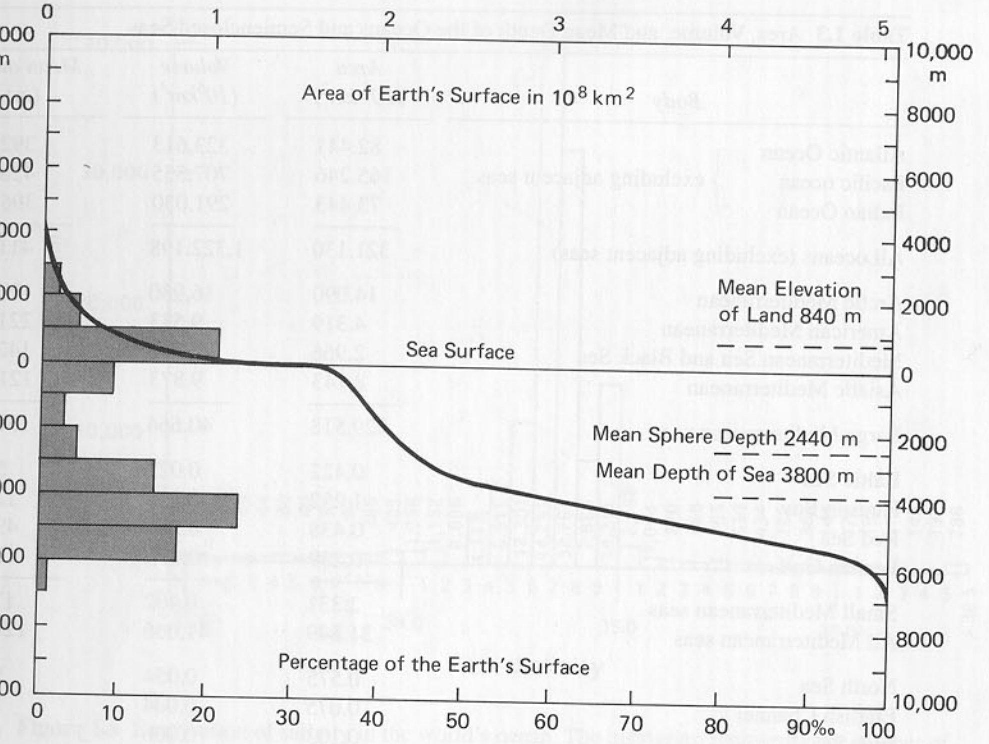


Figure 1.7 Hypsographic curve showing area of earth's solid surface above any given level of elevation or depth. At left in the figure is plotted the frequency distribution of elevations and depths in 1000-m intervals. (After Sverdrup, Johnson, and Fleming, *The Oceans*, Prentice Hall, 1942.)

the different oceans cause variations from this simple picture, but in all oceans one can see evidence of these great circular gyres of wind-driven currents (Figure 1.9).

The way in which the winds drive these major currents is more complicated than the simple assumption that they are a direct product of the drag of the wind on the surface of the ocean. If that were the case, one might expect the strength of these currents and their direction to be closely correlated with the direction and strength of the winds, and they are not; nor are these current gyres symmetrical. The currents on the western sides of the ocean, such as the Gulf Stream, are stronger and deeper than those on the eastern sides of the oceans, such as the Peru or California Current. As discussed in Chapter 6, this intensification of the *western boundary currents* is a result of the rotation and spherical shape of the earth.

One should not interpret ocean current charts, such as Figure 1.9, as meaning that the major ocean currents are both steady and constant. Although there is always a Gulf Stream, the position of the Gulf Stream can vary (Figure 1.10), as can its intensity and the details of its internal structure. Eddies and rings can be spun off (Figure 1.11). In the interior of the ocean, some distance from the major current gyres, the ocean current structure is even more complex. There is as much day-to-day and week-to-week variability in the

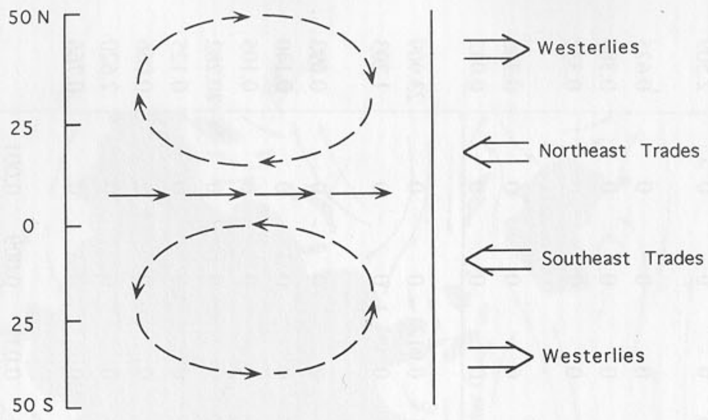


Figure 1.8 The winds apply a clockwise torque in the Northern Hemisphere and a counterclockwise torque in the Southern Hemisphere which set up two current gyres. In the Atlantic and Pacific Oceans, a countercurrent separates the two gyres and is found slightly north of the equator between the Northeast and Southeast Trades of the Northern and Southern Hemispheres.

direction and speed of the surface currents as there is of the surface winds in the interior of the United States.

Thermohaline Circulation. The thermohaline circulation is determined by the density structure of the ocean. Since the density of the water is determined by its temperature and salinity (ignoring compressibility), oceanographers use the term *thermohaline*. Seawater acquires its temperature and salinity characteristics in the shallow surface layers. Its temperature and salinity are determined by processes that take place at the air–sea interface, such as heating by the sun, evaporation, and dilution by rain. Once the seawater leaves the surface layer, temperature and salinity become *conservative properties*; their values can only change by mixing with surrounding water of different temperature and salinity.

A simple picture of the thermohaline circulation is to assume that water mixes along lines of constant density, *isopycnals* (Figure 1.12). One can trace the water's origin,—that is where it was at the surface—by tracking temperature and salinity along isopycnals and noting where water outcrops at the surface. Doing so suggests that the deepest water comes from the most polar latitudes, whereas water of intermediate depths comes from lower, more intermediate latitudes. This deeper water moves and mixes slowly and, once well below the surface, it can take a long time to resurface. Water that upwells in the Pacific may have sunk below the surface off Greenland in the North Atlantic a thousand years earlier. There are a number of exceptions to this concept of slow movement along isopycnals, some of which are discussed in Chapters 7 and 8, but this is a useful beginning to a complicated subject.

Table 1.4 Distribution of Depth Intervals in the World Oceans

Ocean	Depth interval in kilometers 0–0.2	0.2–1	1–2	2–3	3–4	4–5	5–6	6–7	7–8	8–9	9–10	10–11	Percent of world ocean in each ocean
Pacific Ocean	1.631	2.583	3.250	6.856	21.796	34.987	26.884	1.742	0.188	0.063	0.019	0.001	45.919
Asiatic Mediterranean	51.913	9.255	10.433	12.151	6.698	7.780	1.636	0.076	0.058	0	0	0	2.509
Bering Sea	46.443	5.975	7.623	10.330	29.629	0	0	0	0	0	0	0	0.625
Sea of Okhotsk	26.475	39.479	22.383	3.403	8.260	0	0	0	0	0	0	0	0.384
Yellow and East China Seas	81.305	11.427	5.974	1.239	0.055	0	0	0	0	0	0	0	0.332
Sea of Japan	23.498	15.176	19.646	20.096	21.551	0.033	0	0	0	0	0	0	0.280
Gulf of California	46.705	20.848	25.891	6.556	0	0	0	0	0	0	0	0	0.042
Atlantic Ocean	7.025	5.169	4.295	8.590	19.327	32.452	22.326	0.738	0.067	0.012	0	0	23.909
American Mediterranean	23.443	10.674	13.518	15.313	20.796	13.440	2.572	0.193	0.051	0	0	0	1.203
Mediterranean	20.436	22.475	17.413	30.515	8.940	0.221	0	0	0	0	0	0	0.693
Black Sea	34.965	12.587	23.077	29.371	0	0	0	0	0	0	0	0	0.140
Baltic Sea	99.832	0.168	0	0	0	0	0	0	0	0	0	0	0.105
Indian Ocean	3.570	2.685	3.580	10.029	25.259	36.643	16.991	1.241	0.001	0	0	0	20.282
Red Sea	41.454	43.058	14.920	0.568	0	0	0	0	0	0	0	0	0.125
Persian Gulf	100.000	0	0	0	0	0	0	0	0	0	0	0	0.066
Arctic Ocean	40.673	16.539	10.209	13.167	16.580	2.834	0	0	0	0	0	0	2.620
Arctic Mediterranean	69.013	20.454	6.274	4.260	0	0	0	0	0	0	0	0	0.766
Percent of world ocean in each depth interval	73.192	4.423	4.376	8.497	20.944	31.689	21.201	1.232	0.105	0.032	0.009	0.001	

After Menard and Smith, *J. Geophys. Res.*, 71, 1966.

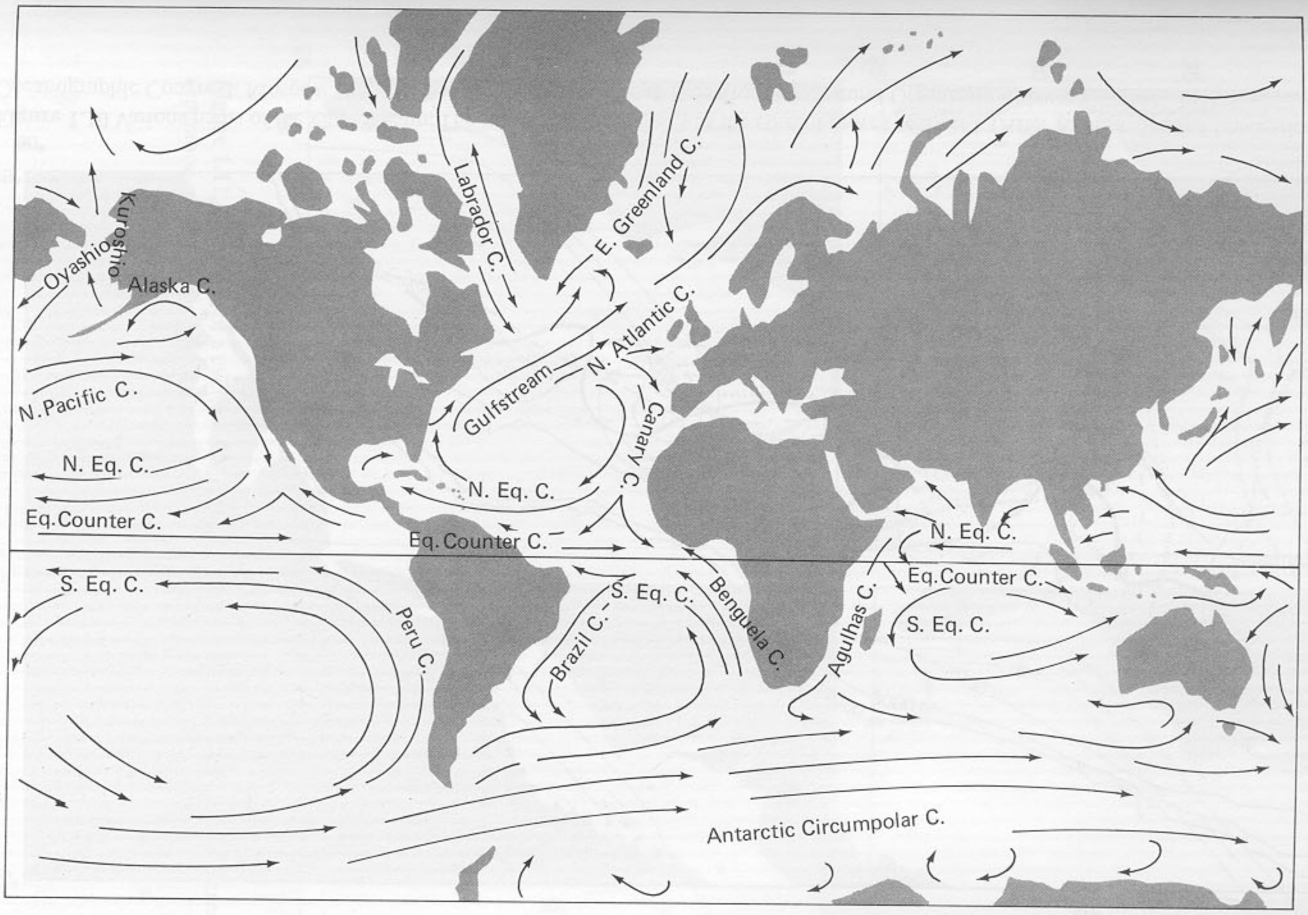


Figure 1.9 The schematic picture of the major surface currents of the world oceans shows some similarity to the simple drawing of Figure 1.8.

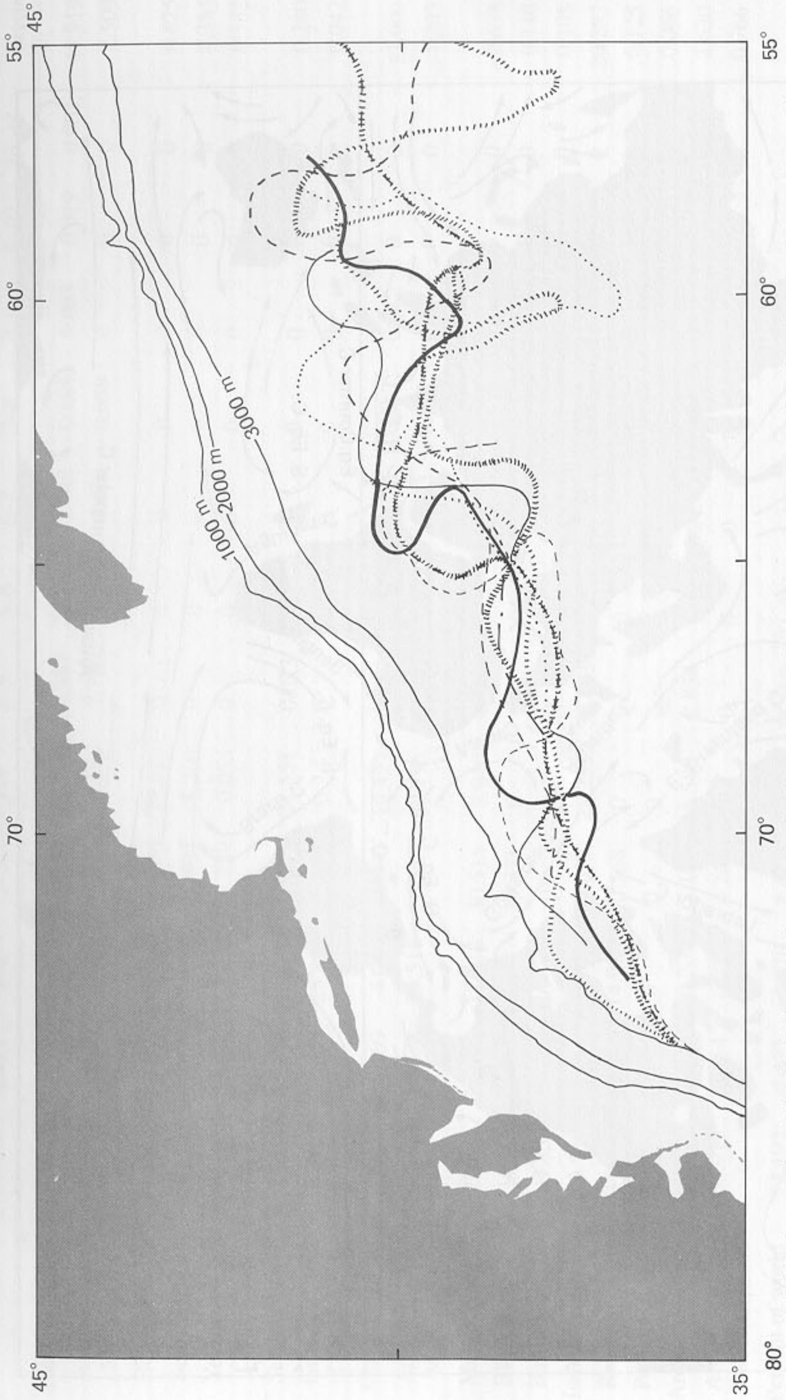


Figure 1.10 Various paths of the Gulf Stream. The meandering increases as the current moves eastward. (After Knauss, Second International Oceanographic Congress, Moscow 1966, United Nations Educational, Scientific and Cultural Organization, 1969.)

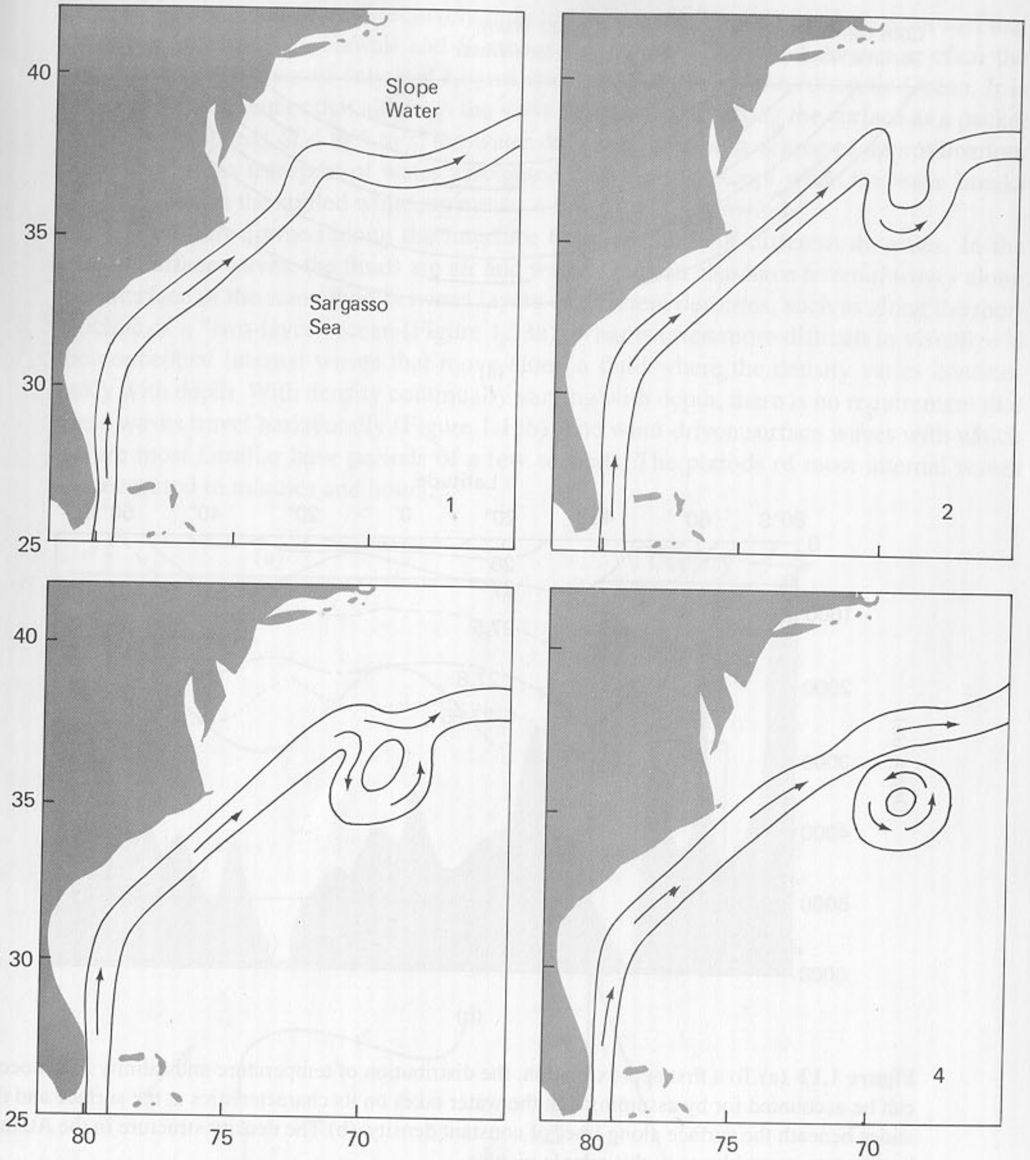


Figure 1.11 Formation of a cyclonic ring. The time between the start of a major meander and the separation of the eddy is typically on the order of 40 days.

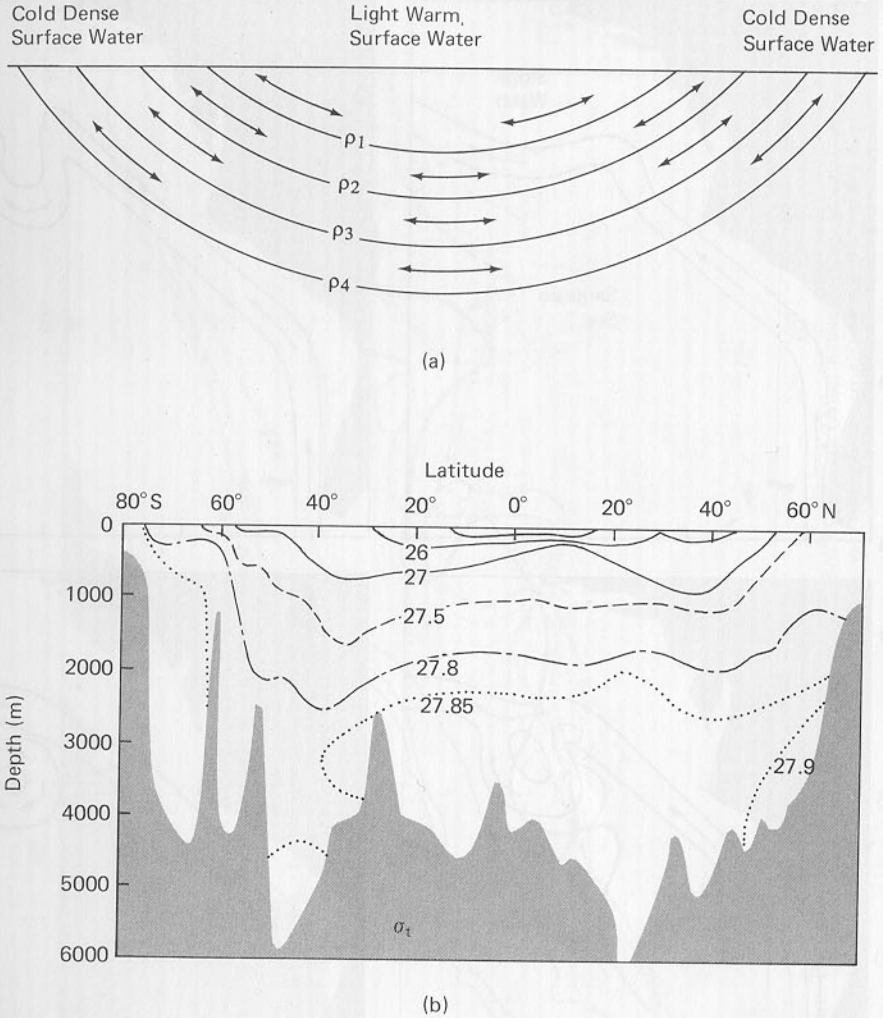


Figure 1.12 (a) To a first approximation, the distribution of temperature and salinity in the ocean can be accounted for by assuming that the water takes on its characteristics at the surface and then slides beneath the surface along lines of constant density. (b) The density structure in the Atlantic bears some resemblance to this simple picture.

The wind blowing on the water surface also generates waves as well as currents. The stronger the wind and the longer the distance, the *fetch,* over which the wind blows, the larger the waves until the wind is blowing the tops off the waves, the waves break, or energy is otherwise being dissipated as fast as the waves are being generated. Under such conditions of equilibrium, the sea is *fully developed.* Once generated, these waves can travel long dis-

tances from their source with relatively little loss of energy. The origin of the giant surf that one sometimes sees in Hawaii and Southern California in July and August is often the major "winter" storms of July and August that are generated in the Antarctic Ocean. It is important to remember that although the wave moves quickly along the surface as a packet of energy at speeds of 4 to 5 m/s, the water does not. To a high degree of approximation, there is no mass transport of water associated with waves, except when the wave breaks and the water in the spilled whitecap moves a few meters downwind.

Waves are formed along the interface between fluids of different densities. In the case of surface waves, the fluids are air and water. One can also have *internal waves* along the interface of the same fluid between layers of different densities, such as along the thermocline of a "two-layer" ocean (Figure 1.13a). What is often more difficult to visualize is the concept of internal waves that move along a fluid where the density varies continuously with depth. With density continually varying with depth, there is no requirement that these waves travel horizontally (Figure 1.13b). The wind-driven surface waves with which we are most familiar have periods of a few seconds. The periods of most internal waves are measured in minutes and hours.

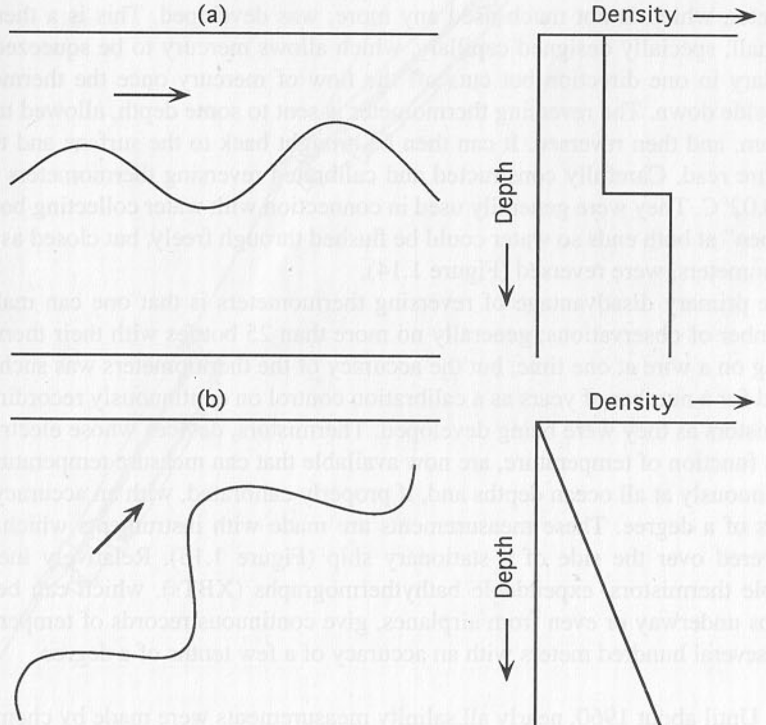


Figure 1.13 (a) Internal gravity waves that move along the density interface of a two layer ocean are analogous to wind-generated surface gravity waves, which move along the two-fluid interface of air and water. (b) In a continuously stratified ocean, there is a gravitational restoring force for waves that move at an angle to the horizontal.

the Ocean Is Observed

Most of our understanding of the ocean comes from observing it and then attempting to apply physical principles to explaining what is observed. Occasionally, a concept is developed from first principles and its existence as an important ocean process is verified by observation. *Double diffusion* is an example (see "Double Diffusion: Salt Fingers" in Chapter 8); but most often in the development of oceanography, it has been the reverse. Measurements are made, and understanding follows. As our ability to observe the ocean improves, so does our understanding. It is important, therefore, to have some knowledge of how the ocean is observed.

Temperature. Until sometime after World War II, the standard of measurement was the mercury thermometer. The problem was how to read the thermometer at a depth of a few tens of meters, let alone at several thousand meters in the deep ocean. There is no practical way to bring the water to the surface without its changing temperature at least a small fraction of a degree; and, of course, there is no practical way to send an observer into the deep ocean to read the thermometers. Since temperature generally decreases with depth, a minimum thermometer was a useful start to solving the problem, but in time the reversing thermometer, which is not much used any more, was developed. This is a thermometer with a small, specially designed capillary, which allows mercury to be squeezed through the capillary in one direction but cuts off the flow of mercury once the thermometer is tipped upside down. The reversing thermometer is sent to some depth, allowed to come to equilibrium, and then reversed. It can then be brought back to the surface and the in situ temperature read. Carefully constructed and calibrated reversing thermometers are accurate to ±0.02°C. They were generally used in connection with water collecting bottles, sent down "open" at both ends so water could be flushed through freely, but closed as they, and the thermometers, were reversed (Figure 1.14).

The primary disadvantage of reversing thermometers is that one can make only a finite number of observations; generally no more than 25 bottles with their thermometers were hung on a wire at one time, but the accuracy of the thermometers was such that they were used for a number of years as a calibration control on continuously recording electrical thermistors as they were being developed. Thermistors, devices whose electrical resistivity is a function of temperature, are now available that can measure temperature rapidly and continuously at all ocean depths and, if properly calibrated, with an accuracy of a few thousands of a degree. These measurements are made with instruments which are carefully lowered over the side of a stationary ship (Figure 1.15). Relatively inexpensive, expendable thermistors, expendable bathythermographs (XBTs), which can be dropped from ships underway or even from airplanes, give continuous records of temperature to a depth of several hundred meters with an accuracy of a few tenths of a degree.

Salinity. Until about 1960, nearly all salinity measurements were made by chemical titration. The ratio of chemical elements in a water parcel varies little from surface to depth and from one ocean basin to another. Thus, titrating for chlorinity, the major negative ion in seawater (Table 1.2), and multiplying by a constant

$$\text{salinity} = 1.80655 \times \text{chlorinity} \tag{1.1}$$

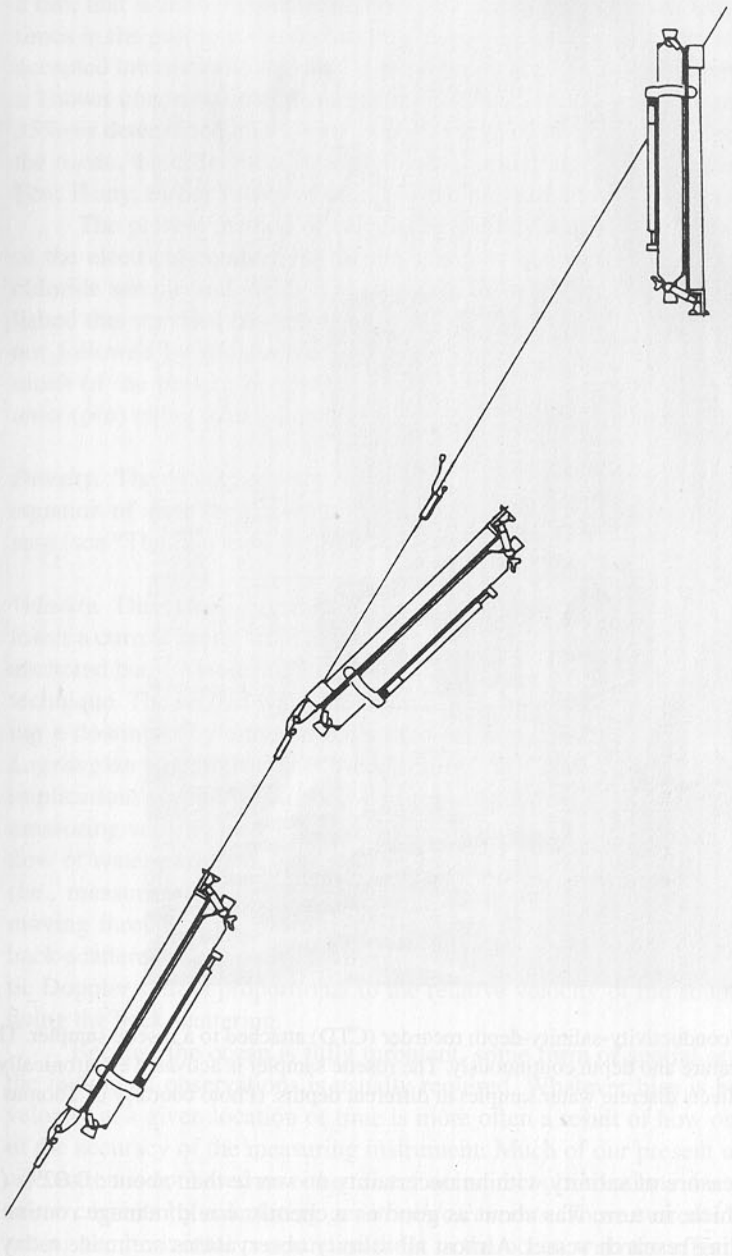


Figure 1.14 For many years, water was collected and temperature measured from a series of bottles and mercury thermometers spaced along a wire lowered into the ocean. A brass weight (the "messenger") slides down the wire, releases the top clamp and allows the bottle to reverse, which in turn releases the next messenger to repeat the process for the next bottle. As the bottles are flipped upside down, valves close capturing water and the flow of mercury in the thermometer is cut off, capturing the water temperature at that depth. As many as 24 bottles would be placed on a wire at one time.

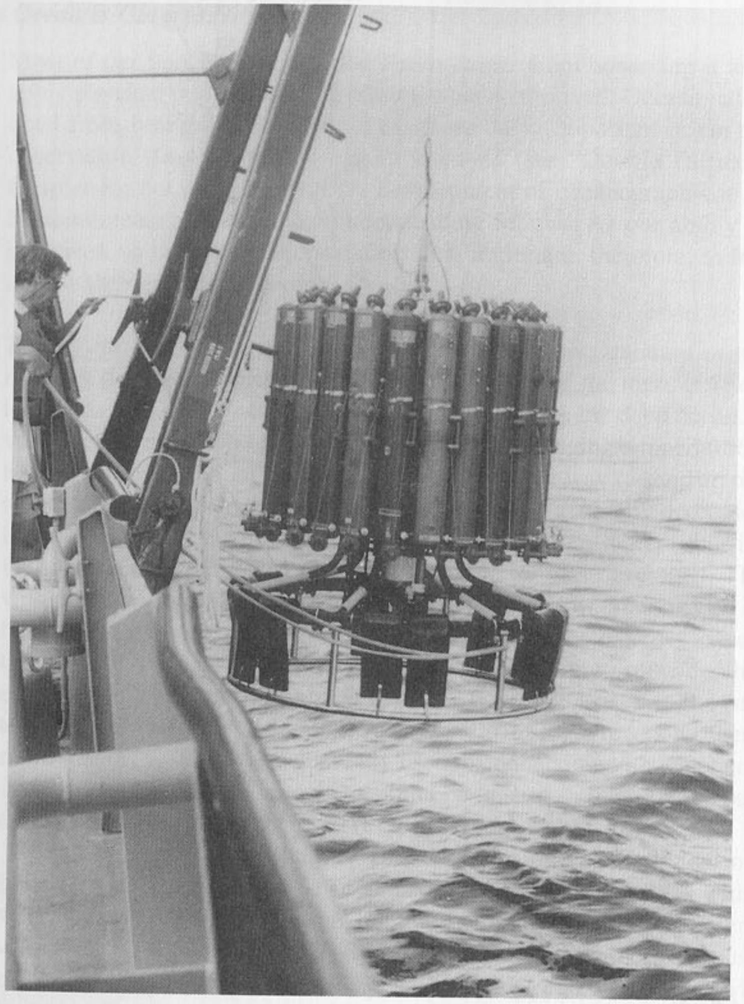


Figure 1.15 A conductivity-salinity-depth recorder (CTD) attached to a rosette sampler. The CTD records temperature and depth continuously. The rosette sampler is activated electronically from the surface and collects discrete water samples at different depths. (Photo courtesy of Thomas H. Rossby)

provides a measure of salinity with an uncertainty no worse than about ±0.02‰ (parts per thousand), which, in turn, was about as good as a chemist could manage routinely on the deck of a rolling research vessel. Almost all salinity observations are made today by measuring the electrical conductivity of seawater. This technique is now the international standard. Present instruments that measure temperature and electrical conductivity in situ can provide an accuracy of ±0.002 psu.

Neither chlorinity titration nor electrical conductivity is a measure of true salinity. Each uses a fundamentally different technique to reach an approximation of salinity. Considerable effort has been made to reconcile the two methods, but some difficulties remain,

a task that is further complicated because different algorithms have been used at different times in the past to convert chlorinity and electrical conductivity to salinity. The presently accepted international standard is to relate the electrical conductivity of seawater to that of a known concentration of potassium chloride, which was chosen to equal a salinity of 35‰ as determined by titration. For the range of salinity and temperature found in 99% of the ocean, the differences in applying the various algorithms are less than ±0.01‰ or psu. Few, if any, earlier values of salinity were measured to that accuracy.

The present method of calculating salinity requires a comparison between the ratio of the electrical conductivity of the seawater sample and that of the standard potassium chloride sample and yields a nondimensional number. The international body that established this standard has recommended that the value calculated be shown as unitless and not followed by (‰), a recommendation that has not been widely accepted. However, much of the present oceanographic literature has substituted the term *practical salinity units* (psu) rather than following the traditional parts per thousand (‰).

Density. The density of seawater is not measured directly. Rather, it is calculated from the equation of state for seawater, based on known values of temperature, salinity, and pressure (see "The Equation of State," Chapter 2).

Velocity. Direct velocity measurements are of two kinds. An example of the first is to lower a current meter from an anchored vessel, or attach one or more current meters to an anchored buoy. Measuring the flow of water by a given point is referred to as the *Eulerian* technique. The second way is to track a particle of water as it moves—for example, tracking a floating buoy either at the surface or at a constant depth. This is referred to as the *Lagrangian* technique. (See "Acceleration" in Chapter 5 for a further discussion of the implications of differences between these two ways.) A number of indirect methods of measuring velocity have evolved. These include estimating the velocity or transport of the flow of water past a point by making use of Faraday's law of induced electromotive force (i.e., measuring the electrical potential generated by an electrical conductor [seawater] moving through the earth's magnetic field). Another is to measure the Doppler shift of back-scattered sound from a transmitter on a ship or a fixed buoy; the amount of the acoustic Doppler shift is proportional to the relative velocity of the sound source and the water doing the back scattering.

Because the ocean is fully turbulent, some form of spatial or temporal averaging of the individual observations is usually required. Whatever bias is built into the calculated velocity at a given location or time is more often a result of how one averages rather than of the accuracy of the measuring instrument. Much of our present understanding of ocean circulation is not based on either direct or indirect observations. Instead, we infer the mean ocean currents by calculating *geostrophic currents*. These currents are determined by balancing the horizontal pressure gradient in the ocean with the effect of the earth's rotation in terms of the Coriolis force (see "Geostrophic Flow" in Chapter 6).

Position. Obviously, one needs to know the geographical position of one's observations of the temperature, salinity, density, and velocity. All of the early observations were made from aboard a research vessel, and many still are. Until the end of World War II, celestial navigation was the only way to position a ship at sea far from the sight of land. Accurate star sights require a good horizon, which meant limiting observations to dawn and dusk. Under ideal circumstances (few clouds and a relatively smooth sea), one's position could be located to within 1 or 2 km twice a day. Knowing one's position between sights required dead reckoning. During daylight the position of the sun provides additional guidance, but not an exact position. Thus, most of the time one's position at sea was not known to better than a few kilometers, and if one had the misfortune to work in regions of strong ocean currents and significant cloud cover, the uncertainty of one's position was on the order of 5 to 10 km and sometimes much more.

After World War II, electronic navigational aids established on shore in the United States and Europe made possible positioning oneself to within 1 or 2 km all of the time as long as the ship was within a few hundred kilometers of the shore stations. Otherwise, one's position was determined as described in the previous paragraph.

The use of satellites as navigational aids began in the 1970s. At that time, satellites provided the equivalent of accurate, cloud-free celestial positions once every few hours, thus reducing significantly the time for dead reckoning between fixes. The current satellite navigational system provides positions 24 hours a day to within a few meters. The position of drifting buoys can also be determined by satellite.

Depth. For observations made beneath the surface, fixing the position of the observation is a three-dimensional problem. The earliest method, and the simplest, was to measure the length of line necessary to reach the bottom or that length connecting the instrument to the ship. On a drifting ship the line is seldom vertical; therefore, without making some allowance for "wire angle," the depth is often overestimated.

Since World War II, the depth of the ocean has been routinely measured by echo sounding. Knowing the velocity of sound in seawater, one can determine the depth by measuring the time for the echo of a sound pulse from a ship to bounce off the bottom and return to the surface. Echo sounding is occasionally used to determine the depth for an instrument not on the bottom, but the positions of most middepth instruments are determined by measuring the hydrostatic pressure:

$$p = \int_0^Z \rho g \, dz \tag{1.2}$$

In the case where density is a constant, the pressure at depth Z is

$$p = \rho g Z \tag{1.3}$$

Because the value of gravity is about 2% less than 10 m/s^2 and the density of seawater is 2 to 4% larger than 1000 kg/m^3, a pressure of 1 decibar (db) is nearly equivalent to 1 m of depth. (A decibar is one tenth of a bar, and a bar is 10^6 dynes/cm^2 or 10^5 Pa.) Instruments that measure pressure to a few decibars are used routinely. If required, instruments are available to provide even greater accuracy. The density of seawater varies with latitude, as does gravity. For those occasions when it is important to know the exact depth, as distinguished from the decibar depth, the table in Appendix 3 can be used. Generally, the true depth in meters is 1 to 2% less than the equivalent decibar depth.

Satellites for Ocean Observing. The number of ways available to physical oceanographers to observe the ocean continues to increase. Sound has been used for measuring the turbulent structure of the ocean on a wide range of spatial scales, and echo sounders placed on the ocean bottom have been used to observe the changing height of the sea surface. No recent technique has provided greater insight into physical oceanographic processes than satellites. Satellites have the virtue of giving wide, almost instantaneous spatial coverage. They have been used for showing patterns of surface temperature, for estimating surface wind speed and direction by measuring the roughness of the ocean surface, and for estimating geostrophic current by measuring the slope of the sea surface. Examples of satellite-derived surface temperature measurements are Plate I and Figure 11.10.

It is to be expected that new instruments will continue to be developed for use on satellites to improve our observational capability. Satellite instrumentation, however, has one significant limitation: All of its observations of the earth depend on electromagnetic measurements, either those reflected from energy beamed to earth or those generated on earth. Unlike the atmosphere, the ocean is virtually opaque to the transmission of electromagnetic energy, whether it be radio waves, X-rays, or sunlight. As a consequence, the ocean observed by satellite is only the surface skin. Instruments aboard satellites do not peer far below the surface.

A Word about Units

The International System of Units (SI units, for Système Internationale), as recommended by the International Association of Physical Sciences of the Ocean, is generally used throughout this text. This system is based on units of meter-kilogram-second. One exception is pressure when referring to pressure as depth. Here the *decibar* has been retained, in part because of its continual wide usage and in part because there is such a close correlation between decibars of pressure and depth of water in meters. In much of the earlier literature and textbooks, including the first edition of this one, the cgs (centimeter-gram-second) notation is used extensively. Appendix 2 is included to help the reader make the necessary conversions from one to another.