

Chapter 4

Oceanographic Variables

"Facts which at first seem improbable will, even on scant explanation, drop the cloak which has hidden them and stand forth in naked and simple beauty." – Galileo Galilei

Now that we have a firm understanding of how our experiment should be structured, it is time to turn our attention to the actual variables we shall be measuring in pursuit of that experimental design. But how do we determine what those variables should be? At the most basic level, we must ask ourselves—what must I measure to reveal the information most pertinent to my hypothesis?

Of course, there are no universal answers—that is a question you must answer for yourself. However, there are some fundamental oceanographic variables that offer the greatest bang for your buck and can at least get you started in the right direction.

Keep in mind that this chapter is not meant to serve as an exhaustive list of only those variables with which to concern yourself. Quite the contrary—it would be impossible to establish in an entire volume, let alone a single chapter, a definitive list of every oceanographic variable of critical import. Rather, this chapter is meant to outline what should be considered a minimalist's view of the most basic variables to consider when formulating your research plan.

Key Concepts

- The Lagrangian frame of reference describes ocean dynamics from the perspective of the fluid parcel itself as it flows through space and time.
- The Eulerian frame of reference describes ocean dynamics from the perspective of the fluid parcel as it moves through a fixed region of interest.
- The measurement of any oceanographic variable must be constrained by appropriate spatial and temporal boundaries, as defined by the investigator.
- There exist certain key oceanographic variables that can be considered essential to virtually all types of investigations of ocean dynamics according to their relevance in the biological, chemical, geological, and physical subdisciplines of oceanography.

Establishing the Frame of Reference

Before deciding on the specific oceanographic variables we will wish to measure in support of our field research, it is necessary that we first consider the appropriate frame of reference for those measurements. This is especially important in the context of aquatic systems because virtually all of our measurements will vary in the three spatial dimensions, as well as in time.

This presents us with one of the fundamental challenges of oceanographic research: Is it possible to simultaneously define four-dimensional variability? And if so, should our measurements be taken from the perspective of the fluid medium itself (within the flow), or should those measurements be taken from a fixed position (exterior to the flow)? What those measurements “mean” will be highly dependent on our frame of reference.

In the context of fluid dynamics, oceanographic observations are typically made from either of two different perspectives (**Figure 4.1**). The **Lagrangian** point of view (POV) seeks to describe fluid motion from the perspective of the fluid parcel itself as it flows through space and time. If we could shrink ourselves down and sit astride a rubber duck floating on the ocean surface, we would be observing the flow from a Lagrangian perspective. By contrast, the **Eulerian** POV demands that we measure the fluid as it moves through a fixed region of interest as time passes. So, if we were sitting on a pier and happened to observe a rubber duck floating by, we would be observing the flow from an Eulerian perspective.

As another example, let us consider an idealized loop current where the stream velocity within the current is much stronger than the flow outside the current in the surrounding water (**Figure 4.2**). With the passage of time, the loop current pinches off, forms a truncated loop, and spawns a small eddy. A drifter caught in the eddy would continue to measure the water properties within the flow; from its Lagrangian perspective, the drifter would not necessarily recognize any differences in the surrounding water properties, or that it was even “trapped” in the eddy. If we were observing these dynamics from a fixed location outside the flow, our Eulerian perspective would make it rather easy to recognize the spatial evolution of the flow with the passage of time. In this case, the data that describe the same oceanographic feature (eddy formation) could be analyzed in very different ways depending on the frame of reference.

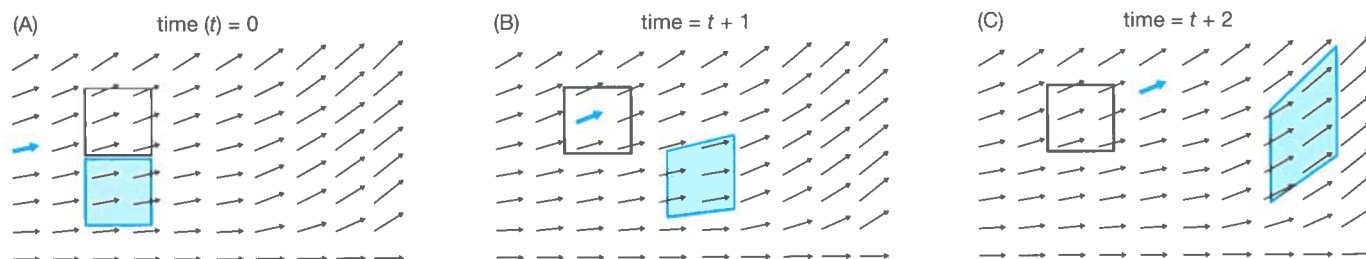


Figure 4.1 A stylized view of water advection, where the length of the arrows represents the relative speed of the current, and the direction of the arrows represents the direction of water movement. In the Lagrangian view (*shaded box*), the region of interest (ROI) travels with the water and is monitored as it evolves over time. In the Eulerian view (*unshaded box*), the ROI remains fixed in a single location and the water is monitored as it flows into and then departs the ROI. Note that at time $t = 0$ (A), the Lagrangian and Eulerian views are indistinguishable from each other, and neither has experienced the discrete parcel of water located upstream from the flow (*blue arrow*). At some future time (B), the Eulerian view experiences the parcel of water (*blue arrow*) as it enters the ROI, while the area within the Lagrangian view is advected downstream. Even further into the future (C), the discrete water parcel (*blue arrow*) exits the Eulerian view and can no longer be measured within the ROI. Of course, the area within the Lagrangian view was never able to witness the passage of that particular water parcel, and has been deformed and advected further downstream by the flow. (Adapted from Janicke H & Scheuermann G, 1981)

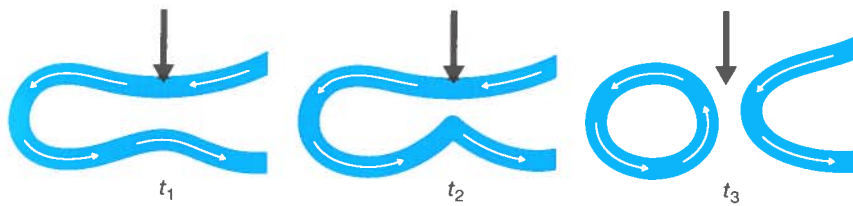


Figure 4.2 In the Lagrangian view, the properties of a water parcel *within the flow* (white arrows) will not seem to change from time t_1 to t_3 . However, from the Eulerian perspective, the properties of the water parcel will seem quite different from t_1 to t_3 , when viewed from a fixed location *outside the flow* (black arrow).

In practice, it is generally much easier to conduct oceanographic studies from an Eulerian perspective. Whether measurements are collected from a coastal landmark, a moored buoy, a research vessel, or from a satellite, the data can be easily related to a latitude–longitude–depth coordinate system—an Eulerian POV. To conduct a truly Lagrangian study, the instrument would be required to maintain its position within the evolving water mass, continuously recording data over the course of its travels. Although such drifters do exist, there are significant technical and financial challenges associated with their deployment, use, and recovery.

If the researcher is truly interested in the evolution of **discrete** water masses over time, it may be possible to perform a pseudo-Lagrangian study by simply adding more stations upstream or downstream of the main region of interest and increasing the sampling interval. As long as the same measurements are being recorded at all stations and at regular time intervals, the continuous evolution of those water masses will be captured in the “snap-shot” measurements recorded at each of the fixed locations over time.

Defining Spatial and Temporal Boundaries

Once you have established your preferred frame of reference, it will be necessary to define the boundaries of your study in space and time. In nearly all oceanographic applications, fixed positions on the Earth’s surface are defined using the two-dimensional coordinate system that defines the **easting** direction in degrees longitude and the **northing** direction in degrees latitude (**Figure 4.3**). For ancient cartographers and navigators, determining one’s exact position on the face of the Earth (with nothing but a compass in hand) was no easy task. With the advent of **GPS** technology, obtaining high-precision global positioning has now become routine.

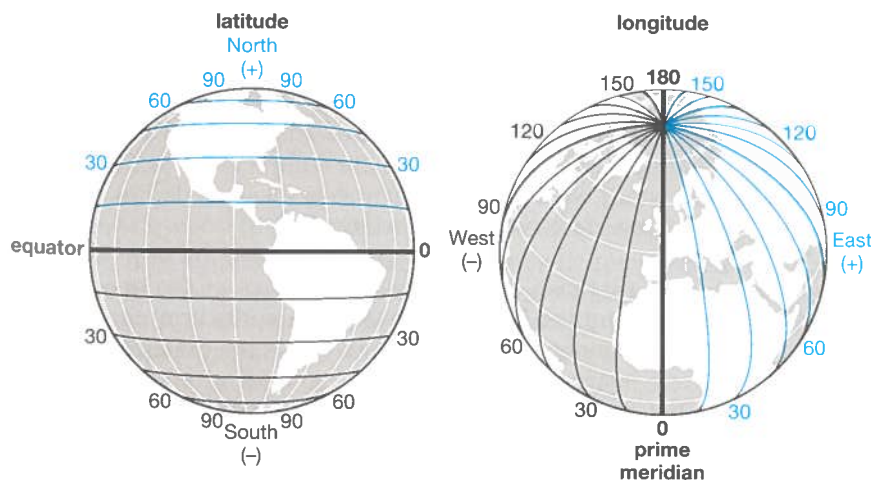


Figure 4.3 The precise location of any object on the Earth can be described using a Cartesian coordinate system, where latitude represents the y -coordinate (0 to 90° north or south of the equator) and the longitude represents the x -coordinate (0 to 180° east or west of the prime meridian).

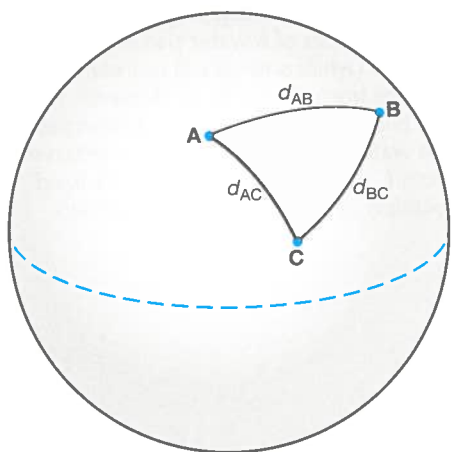


Figure 4.4 Although the latitude and longitude coordinates at points A, B, C can be easily determined using GPS technology, it may be more meaningful to relate these points to each other as a function of their relative distances (d) from each other. For distance calculations performed on a curved plane, we must use the Haversine functions.

Distance Measures are a Simple but Critical Way to Define Spatial Relationships

In some cases, it may be more useful to define the horizontal dimension not simply as latitude and longitude coordinates, but as distances from a fixed reference point. For example, if you wished to collect water samples at regular 20-m intervals offshore from the coastline, the latitude and longitude coordinates for all of your sampling sites would differ by only a few arc seconds. In this case, it might make more sense to define the latitude and longitude coordinates of the initial reference point at the coastline, and describe your sampling stations according to their relative distances from that point.

Depending on the sophistication of the surveying equipment being used, horizontal distances of 500 m or less can usually be measured directly. Over moderate distances (1 to 5 km), it becomes increasingly difficult to survey distances with sufficient accuracy and precision. When distances exceed 5 km, the curvature of the Earth becomes significant enough that we can no longer calculate distances according to planar geometry.

To calculate the distance between two points on a curved surface (Figure 4.4), we must use the **Haversine functions** (see Equations 4.1–4.4). Using a volumetric mean radius r of 6,371.0 km for the Earth, we can calculate the distance d between two points on the Earth's surface as

$$\Delta lon = lon_2 - lon_1 \quad (4.1)$$

$$\Delta lat = lat_2 - lat_1 \quad (4.2)$$

$$a = \frac{[1 - \cos(\Delta lat)]}{2} + \left\{ \cos(lat_1) \cdot \cos(lat_2) \cdot \frac{[1 - \cos(\Delta lon)]}{2} \right\} \quad (4.3)$$

$$d = r \cdot \left[2 \tan^{-1} \left(\frac{\sqrt{a}}{\sqrt{1-a}} \right) \right] \quad (4.4)$$



A HAVERSINE EXAMPLE

If we wanted to calculate the distance between the latitude and longitude coordinates (+45.00, –100.00) and (+35.00, –98.00), our critical values for the Haversine functions are

$$\Delta lon = 0.034907$$

$$\Delta lat = -0.174533$$

$$lat_1 = 0.785398$$

$$lon_1 = -1.745329$$

$$lat_2 = 0.610865$$

$$lon_2 = -1.710423$$

$$a = 0.007773$$

$$d = 1,125 \text{ km}$$

Note that all latitude and longitude coordinates were converted from decimal degrees to radians using the conversion factor $\pi/180$ radians/degree.

where all latitude and longitude coordinates must first be converted to **radians**, yielding the computed distance d in kilometers.

Depth Measures are Used to Extend Spatial Relationships into the Vertical Dimension

In aquatic systems, it is not sufficient to merely define our horizontal dimension; we must also define the vertical dimension as well. But measuring discrete ocean depths is far trickier in practice than it would at first seem. To accurately measure ocean depths, we must first establish a fixed reference point in the vertical to represent our “zero depth.” This is where our troubles begin, because the surface of the ocean is never uniform—it is constantly being deformed by physical forces beyond our control (such as the tides, or swells caused by wind). So how do we accurately measure depths from the ocean surface if that surface is constantly changing?

One solution is to establish your frame of reference at the ocean floor, measuring heights from the bottom rather than depths from the surface. In many physical oceanography applications, this is the preferred method (especially if your instruments are designed to float at various heights, tethered to the bottom). If it is necessary to measure depths from the surface, the best solution is to use a submersible pressure sensor, such as a

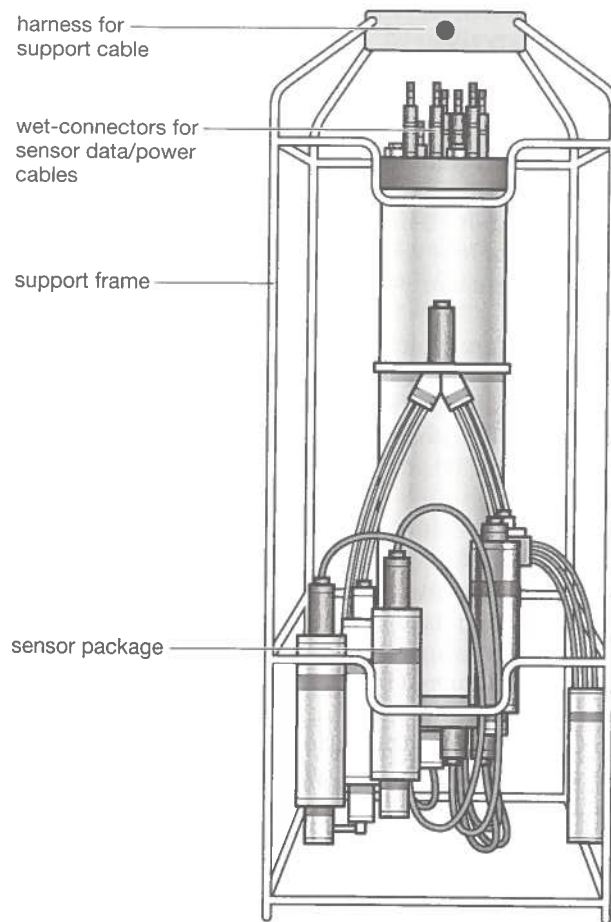


Figure 4.5 An integrated conductivity-temperature-depth (CTD) Recorder (SBE 9 model). (Courtesy of Seabird Electronics, SBE 911plus model.)

conductivity-temperature-depth (CTD) recorder (**Figure 4.5**), to continuously record the **hydrostatic pressure** of the water column above the device as it is lowered. With such devices, any changes in hydrostatic pressure can be instantaneously correlated with the density and height of the overlying water column that is causing those pressure changes. In this way, the exact height of the water column (from the submerged sensor to the ocean surface) can be measured and defined as the *in situ* depth.

Although it is possible to measure water depths from the surface without a CTD, such low-tech methods usually require the use of a marked and weighted line that is dropped to the appropriate depth (an inherently imprecise solution). Such measurements can be improved by referencing the measurements against the **mean sea level** (MSL), but such efforts require the investigator to correct for any influences caused by currents, waves, and tides, as they will cause the local sea level to differ (sometimes significantly) from MSL.

Time is Required to Define the Succession of Events

With the three-dimensional spatial scale defined, any measurement taken at a specific location can only be represented as an instantaneous value, unless we are conducting those measurements at regular time intervals. Defining the temporal dimension of our variables is absolutely critical to success, as we must adhere to time scales that are relevant to the phenomena we wish to capture in our data. For example, if we were interested in studying the tidal cycle at various coastal locations, we might have a very

difficult time making sense of our data if we decided to take our measurements at 15-hour intervals.

It is also important to consider how time should be scaled. For most long-term investigations, data are usually depicted first by the calendar year, and then by the calendar day (January 1 = Day 1, December 31 = day 365 for non-leap years). To compensate for leap years (or if the study has a significant seasonal component to it), it may be beneficial to convert the calendar day to 2π radians (360°), thereby relating the calendar day to the Earth's relative "completion" of its annual orbit around the Sun.

Shorter time intervals are typically denoted using the 24-hour military-style clock, as HH:MM:SS. Using this convention, the time of day is counted from midnight to midnight and is divided into 24 hours. As an example, 7:00 a.m. and 6:30 p.m. would be cited as 07:00:00 and 18:30:00, respectively (with midnight returning to 00:00:00).

Because time of day is relative to longitude, the global timekeeping convention usually considers 0° longitude (the **prime meridian**) as the global "time zero." Because Greenwich, England, is located on the prime meridian, the time designation used is sometimes called Greenwich Mean Time (GMT). In more recent years, the GMT designation has been replaced by Universal Time Coordinated (UTC), but they are synonymous.

Using the UTC convention, the various time zones across the Earth are calculated as positive or negative offsets from UTC (**Figure 4.6**). Based on your specific longitude, it would be necessary for you to determine the appropriate UTC offset if you wished to cite your time-stamped data relative to UTC. For example, the UTC offset for Puerto Rico is -4.0 hours. If we were lucky enough to be performing field research off Puerto Rico and took our measurement at 11:34:27 local time, that time-stamp is -4.0 hours from UTC. Therefore, we would simply have to add 4.0 hours to cite our data consistent with UTC (collected at 15:34:27 UTC, in fact).

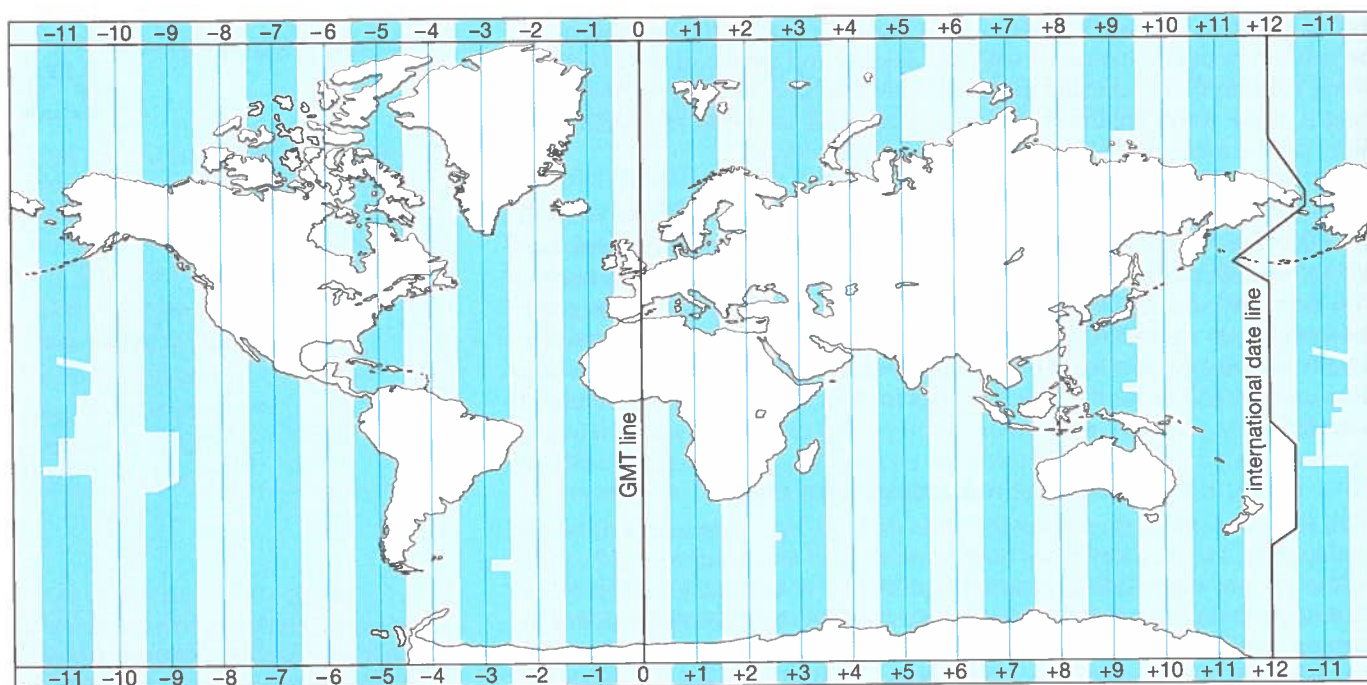


Figure 4.6 UTC time zones of the world.

Physical Variables for Consideration and the Essential Measures of Ocean Physics

When we speak in terms of physical oceanography, we are essentially attempting to describe the physical structure of the ocean and how the ocean obeys the laws of motion, the laws of thermodynamics, and the laws of conservation of mass and energy. When these fundamental concepts of physics are considered within the context of ocean science, we are able to elucidate the variables that more clearly define those phenomena of special interest, such as fluid dynamics (like waves, currents, and tides) as well as the propagation of light and sound within the ocean.

The physical properties of the ocean that are most important to fluid dynamics and ocean structure are the measurements of a water parcel's mass m , temperature T , and pressure P . With regard to seawater, mass cannot be considered solely from a physical perspective; in fact, the mass of seawater is intimately related to its salinity S —a chemical property. Remarkably, measurements of water mass, temperature, and pressure can be used by a clever oceanographer to derive (or estimate) many other physical variables of relevance, as we will soon discuss.

Mass is the Most Important Way to Quantify Existence

When we think of an object's mass, most of us (mistakenly) think in terms of the object's weight; as if the two were perfectly synonymous. Technically, an object's mass is a measure of the quantity of **matter** that comprises that object. One of the most fundamental principles in the physical sciences is that matter can be neither created nor destroyed. In other words, the amount of matter within an object must either remain unchanged, or must be 100% accounted for—this is the law of conservation of mass. Thus, the amount of matter will be absolute, whereas the “weight” of that matter will always be relative to the balance of forces exerted on that mass while it sits on the scale. Imagine if we stepped onto a scale while we were shoulder deep in a swimming pool—the buoyancy of our submerged bodies would counteract the pull of Earth's gravity, and the scale would read as if we were much lighter than we really are (but our true mass would remain unchanged).

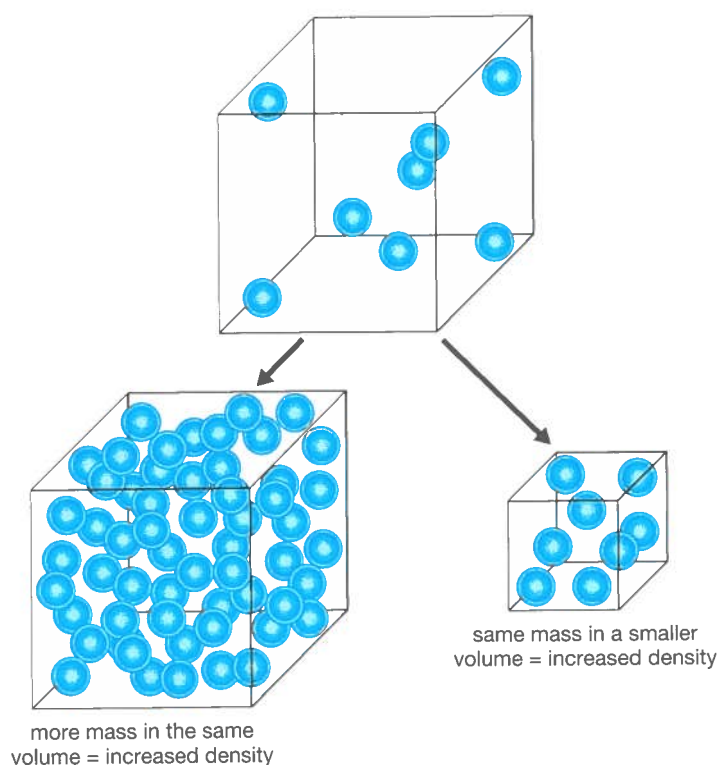
In most cases, measuring the mass of a solid object is so elementary that we don't even think about it. After all, it is not difficult to draw the boundary that separates an 8-kg bowling ball from the rest of the natural world. It's an entirely different challenge for us to define the physical boundaries that separate an 8-kg mass of fluid from the rest of the ocean. How do we conceptually “draw the line” between 8 kg of ocean, or perhaps 8 kg of atmospheric gas?

When dealing with the masses of liquids or gases, it is necessary for us to also define the three-dimensional space that “contains” the fluid (**Figure 4.7**). This is why we use **density** (ρ , in units of g cm^{-3} , or kg m^{-3}) as the preferred expression of fluid mass in oceanography. Thus, the measure of a fluid's density is simply the quantity of matter (its mass) that simultaneously occupies a certain amount of space (its volume). Density (ρ) can be easily determined by calculating the mass-to-volume ratio; or we can express the density of a solid or fluid by the **specific volume** α it occupies:

$$\rho = \frac{m}{V}; \quad \alpha = \frac{1}{\rho} \quad (4.5)$$

For example, a fluid sample with a density ρ of $1012.35 \text{ kg m}^{-3}$ would also have a specific volume α of $9.87801 \times 10^{-4} \text{ m}^3 \text{ kg}^{-1}$.

Figure 4.7 The density of a solid (or fluid) is defined by the amount of mass contained within a specific volume. However, it is important to note that density is affected either by changes to the amount of mass or by changes to the volume it occupies. For example, if more mass is packed into the same volume (*lower left*), density will increase. Likewise, if the same amount of mass is squeezed into a smaller volume (*lower right*), density will increase as well.



Because it is necessary for us to define a fluid's mass in terms of the volume it occupies, we must consider the total mass contained within that volume. That makes things a bit more difficult for us because we can rarely expect to encounter a pure fluid in nature. In the "simple" case of ocean water, we know that any volume of fluid we collect will contain a certain mass of water, several different species of dissolved salts, and several different species of dissolved gases (all with different atomic masses and in various concentrations).

Since the density of a fluid is defined by both its mass and volume, changes to either of those will result in a different measure of overall density. In the context of oceanography, changes in salinity can have a dramatic effect on the overall mass of the water parcel, thereby changing its density. Changes in temperature and pressure can have considerable effects instead on the volume of the water parcel, also resulting in a modified measurement of density.

As you can see, this makes an accurate measurement of density completely dependent on our ability to assess the mass m , temperature T , and pressure P simultaneously. But before we get too ahead of ourselves, let's take a closer look at the effects of temperature and pressure—we'll rejoin our discussion of density a bit later.

Temperature Affects the Physical Properties of All Matter

Most of us are so used to measuring temperatures that we rarely give it much thought. In the context of oceanography, the simple task of measuring the ambient temperature (whether at the surface or at discrete depths) is really a measurement of the *in situ* temperature T . When we are dealing with relatively shallow water depths, it is usually sufficient to measure T .

However, because seawater is slightly compressible, any water parcel at great depth will be compressed into a smaller volume. If that same water parcel

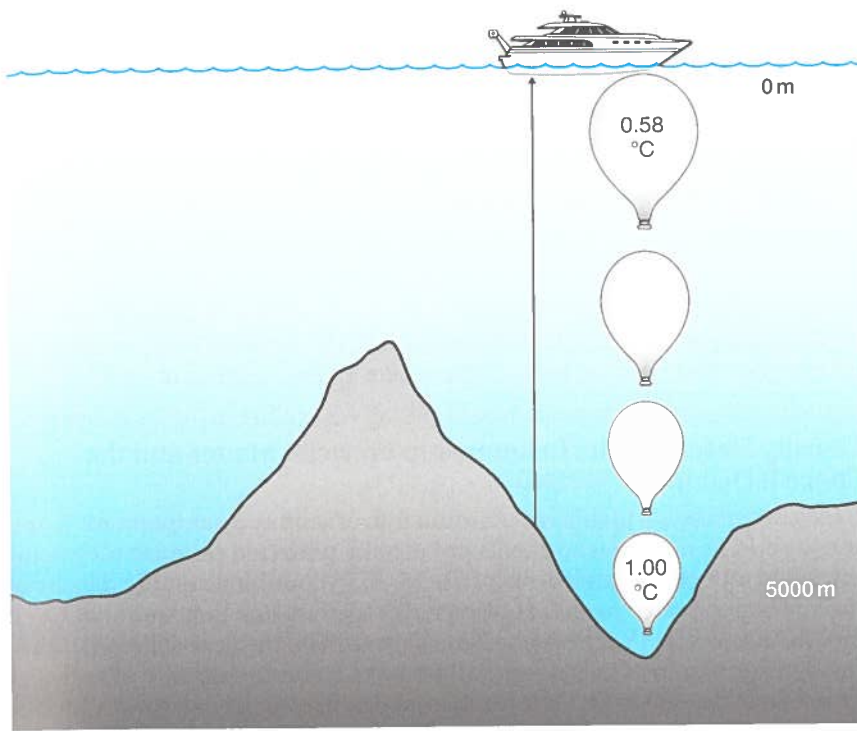


Figure 4.8 The *in situ* temperature of a parcel of water 5000 m deep may be 1.00°C at depth, but if this parcel were raised to sea level, the relief in hydrostatic pressure would result in the adiabatic expansion of the water parcel, causing it to cool to 0.58°C (its potential temperature, or θ).

were raised to the surface without gaining or losing heat to the surrounding water, the alleviation of pressure would cause the water parcel to slightly expand and therefore cool (**Figure 4.8**). If we were to measure the theoretical temperature of this deep-ocean water parcel after raising it to the surface, we would measure it as the parcel's potential temperature (θ). Potential temperatures are extremely useful measures when comparing water samples from different depths, or when water masses are transported over significant vertical distances, because θ compensates for any and all compressibility effects (whereas measures of *in situ* temperature do not).

Regardless of whether temperatures are measured *in situ* (T) or as θ , the preferred scale of measurement is in decimal degrees Celsius (°C). In oceanography, seldom are temperatures measured in degrees Fahrenheit (°F); however, the two are easily interchangeable using

$$^{\circ}\text{F} = \left(\frac{9}{5} \cdot ^{\circ}\text{C} \right) + 32 \quad (4.6)$$

$$^{\circ}\text{C} = \left(\frac{5}{9} \right) \cdot (^{\circ}\text{F} - 32) \quad (4.7)$$

Occasionally, it may be necessary to convert to kelvins (K), a measure of absolute temperature. The Kelvin scale is simply defined as

$$\text{K} = ^{\circ}\text{C} + 273.15 \quad (4.8)$$

Hydrostatic Pressure is Used to Describe the Potential Energy within a Fluid

In most oceanographic applications, pressure (P) is measured *in situ* and refers only to hydrostatic pressure and excludes the contributions of atmospheric pressure. In most cases, the variations in hydrostatic pressure far exceed the subtle variations in atmospheric pressure, so we can

assume that standard atmospheric pressure is essentially a constant (1 atm = 1013.25 mbar, or 101.325 kPa). Thus, by removing the effects of atmospheric pressure, $P = 0$ kPa at the ocean surface.

The best course of action is to measure the actual hydrostatic pressure P_h *in situ*. However, if it is necessary to estimate P , its simplest algebraic form is

$$P_h = \rho g d \quad (4.9)$$

where ρ specifically represents the density of the overlying water column (in kg m^{-3}), g is the mean constant of gravitational acceleration (9.8 m s^{-2}), and d is the discrete depth of measurement (in meters). Hydrostatic pressure (P_h) is calculated in units of kPa, where $1 \text{ kPa} = 1000 \text{ kg m}^{-1} \text{ s}^{-2}$.

Density Describes the Relationship between Matter and the Space it Occupies

Because density is a highly variable function of salinity S , temperature T , and pressure P , *in situ* measurements are usually preferred because they automatically reflect the true effects of the (S , T , P) conditions on the density of the measured water parcel. However, this also means that we must cite ρ as a function of its (S , T , P) conditions where the data were collected. Since the density of pure, fresh water is 1000 kg m^{-3} and the density of seawater rarely exceeds 1060 kg m^{-3} , the total variability in ρ , for all the world's oceans, will occur only in the last two figures. To focus our attention where it really belongs (on those last two figures), many oceanographers convert ρ (S , T , P) to **sigma** (σ):

$$\sigma(S, T, P) = [\rho(S, T, P) - 1000 \text{ kg m}^{-3}] \quad (4.10)$$

If we cannot measure *in situ* density, we must instead estimate ρ according to its unique (S , T , P) conditions. For most applications where the accuracy of ρ can be $\pm 0.5 \text{ kg m}^{-3}$, the simple linear equation of state should be sufficient:

$$\rho(S, T, P) = [\bar{a}(T - T_0) + \bar{b}(S - S_0) + \bar{k}P] + \rho_0 \quad (4.11)$$

where $\bar{a} = -0.15 \text{ kg m}^{-3} \text{ per } ^\circ\text{C}$, $T_0 = 10^\circ\text{C}$, $\bar{b} = 0.78 \text{ kg m}^{-3} \text{ per } \text{‰ salinity}$, $S_0 = 35\text{‰ salinity}$, $\bar{k} = 4.5 \times 10^{-3} \text{ kg m}^{-3} \text{ per decibar}$, and $\rho_0 = 1027 \text{ kg m}^{-3}$. When a more precise estimate of $\rho(S, T, P)$ is needed, it is preferable to use the 1980 International Equation of State (IES 80) of Seawater (**Technical Box 4.1**).

Supplemental Measures of Ocean Physics

Within the context of physical oceanography, measures of *in situ* temperature, density, and hydrostatic pressure are so critical, in so many different applications, that they should be collected as part of any research effort. Keep in mind that as important as these measurements are, it may be necessary to expand your list of physical parameters to include those variables that have a more direct influence on the fluid dynamics of the ocean.

Flow Fields (and Currents) are Used to Describe both the Speed and Direction of Water Movement

Because fluids are free to flow in three dimensions, it is often necessary to determine the speed and direction of these flows (as distinct **currents**) in three dimensions. In practice, this is done using a directional current meter (**Figure 4.9**), a submersible device that can measure current speeds in three different directions simultaneously: the easting (u), the northing (v), and the

TECHNICAL BOX 4.1

1980 International Equation of State (IES 80) of Seawater

The most recent (and most precise) accepted method for estimating the density of sea water at any given salinity S (in ‰), temperature T (in °C), and hydrostatic pressure P (in bars) is defined by the IES 80 as the function

$$\rho(S, T, P) = \frac{\rho(S, T, 0)}{\left(1 - \left[\frac{P}{K(S, T, P)}\right]\right)} \quad (4.12)$$

where $\rho(S, T, 0)$ represents the density of seawater at the surface ($P = 0$), such that

$$\begin{aligned} \rho(S, T, 0) = & 999.842594 \\ & + (T \cdot 6.793952 \times 10^{-2}) - (T^2 \cdot 9.095290 \times 10^{-3}) \\ & + (T^3 \cdot 1.001685 \times 10^{-4}) - (T^4 \cdot 1.120083 \times 10^{-6}) \\ & + (T^5 \cdot 6.536332 \times 10^{-9}) + (S \cdot 8.24493 \times 10^{-1}) \\ & - (T \cdot S \cdot 4.0899 \times 10^{-3}) + (T^2 \cdot S \cdot 7.6438 \times 10^{-5}) \\ & - (T^3 \cdot S \cdot 8.2467 \times 10^{-7}) + (T^4 \cdot S \cdot 5.3875 \times 10^{-9}) \\ & - (S^{3/2} \cdot 5.72466 \times 10^{-3}) + (T \cdot S^{3/2} \cdot 1.0227 \times 10^{-4}) \\ & - (T^2 \cdot S^{3/2} \cdot 1.6546 \times 10^{-6}) + (S^2 \cdot 4.8314 \times 10^{-4}) \end{aligned} \quad (4.13)$$

$K(S, T, P)$ is a complex regression equation, determined experimentally, that represents the secant bulk modulus, mathematically defined by the function

$$\begin{aligned} K(S, T, P) = & 19,652.21 \\ & + (T \cdot 148.4206) - (T^2 \cdot 2.327105) \\ & + (T^3 \cdot 1.360477 \times 10^{-2}) - (T^4 \cdot 5.155288 \times 10^{-5}) \\ & + (P \cdot 3.239908) + (T \cdot P \cdot 1.43713 \times 10^{-3}) \\ & + (T^2 \cdot P \cdot 1.16092 \times 10^{-4}) - (T^3 \cdot P \cdot 5.77905 \times 10^{-7}) \\ & + (P^2 \cdot 8.50935 \times 10^{-5}) - (T \cdot P^2 \cdot 6.12293 \times 10^{-6}) \\ & + (T^2 \cdot P^2 \cdot 5.2787 \times 10^{-8}) + (S \cdot 54.6746) \\ & - (T \cdot S \cdot 6.03459 \times 10^{-1}) + (T^2 \cdot S \cdot 1.09987 \times 10^{-2}) \\ & - (T^3 \cdot S \cdot 6.1670 \times 10^{-5}) + (S^{3/2} \cdot 7.944 \times 10^{-2}) \\ & + (T \cdot S^{3/2} \cdot 1.6483 \times 10^{-2}) - (T^2 \cdot S^{3/2} \cdot 5.3009 \times 10^{-4}) \\ & + (P \cdot S \cdot 2.2838 \times 10^{-3}) - (T \cdot P \cdot S \cdot 1.0981 \times 10^{-5}) \\ & - (T^2 \cdot P \cdot S \cdot 1.6078 \times 10^{-6}) + (P \cdot S^{3/2} \cdot 1.91075 \times 10^{-4}) \\ & - (P^2 \cdot S \cdot 9.9348 \times 10^{-7}) + (T \cdot P^2 \cdot S \cdot 2.0816 \times 10^{-8}) \\ & + (T^2 \cdot P^2 \cdot S \cdot 9.1697 \times 10^{-10}) \end{aligned} \quad (4.14)$$

Note that ρ is given in units of kg m^{-3} and the secant bulk modulus K is given in units of bars.

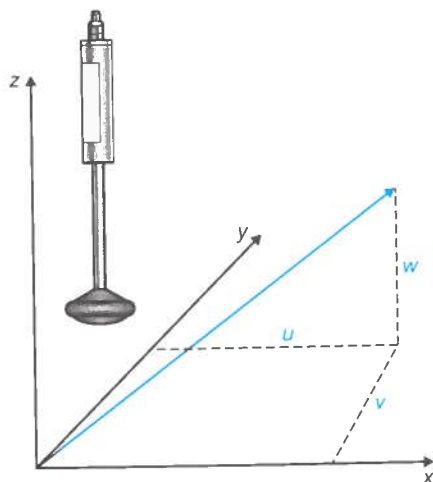


Figure 4.9 An example of a submersible three-dimensional current meter, which can simultaneously measure the (u, v, w) vector components of any flow.

vertical (w) vector components of flow. Since the movement of water within the ocean is described in terms of both the speed and direction of any given current, it is often critical to measure the evolution of three-dimensional flow fields as they might affect particle **advection**, mixing, and the overall stability of the water column.

Wave Geometry Is Used to Discern Wave Patterns in a Complex Ocean

For most of us, it is impossible to think about the ocean without also thinking about ocean waves. Whether we observe them as breaking waves on the shore, or as rising and falling swells causing our vessel to bob like a cork on the ocean surface, ocean waves are an integral part of water movement. When we seek to define wave dynamics, it is most helpful for us to consider the ideal case of an individual wave, propagating in a specific direction, exhibiting the classic sinusoidal wave geometry (**Figure 4.10**).

If we were fortunate enough to encounter something so simple as a single wave in the ocean, it would be a relatively easy task to measure its length λ or its height h . If we had a stopwatch, we could also measure the wave's frequency f and period T with very little difficulty. Unfortunately, the real ocean is never so agreeable—in fact, what we observe in the ocean is a complex mixture of many different waves, of many different wave geometries, that all “arrive” at our location according to their different wave periods and frequencies. Because of this complexity, it is often necessary to describe the sea state according to the **significant wave height** H_s , which is defined as the average wave height of the highest third of all wave heights (**Figure 4.11**).

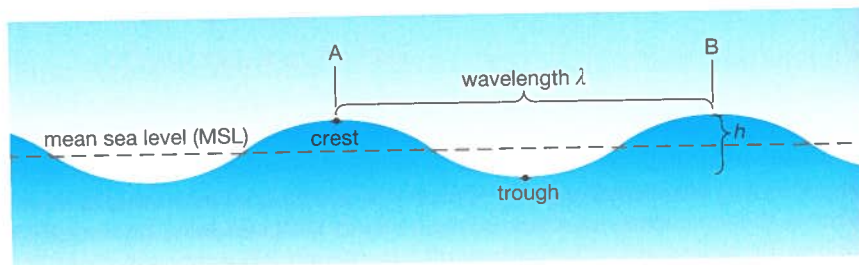
Tides Represent the Influence of Heavenly Bodies on Ocean Dynamics

Among the most significant of all waves within the ocean are our tides, generated primarily by the gravitational forces between the Earth, Sun, and Moon. Because of the stability of the Earth–Sun and Earth–Moon orbits, our tides produce tidal amplitudes (tidal wave heights) and currents that ebb and flow with remarkable predictability. Because of this regularity, the Earth's tides exert a tremendous influence on virtually every aspect of ocean science. As a practical matter, field researchers must always consider how these tidal influences might complicate the design of their field experiments and/or the acquisition of field data.

Secchi Depth Is Used as a Simple Way to Define the Optical Properties of Water

Although the penetration of light in the ocean is most relevant in the context of photosynthesis (see the section “Biological Variables”), the transmission and attenuation of light is largely a physical phenomenon. In most cases, the effective depth of penetration for natural sunlight can be estimated quite easily using a Secchi disk (**Figure 4.12**). Using this simple methodology, a weighted disk is attached to a marked line and lowered until the disk first

Figure 4.10 An idealized ocean wave, indicating the defining elements of its sinusoidal geometry, including the wave crest, trough, length (λ), and height (h). The amount of time required for the wave crest to travel from point A to point B (that is, a single wavelength) is defined as the wave period T . The number of wave crests passing point A every second is defined as the wave frequency f .



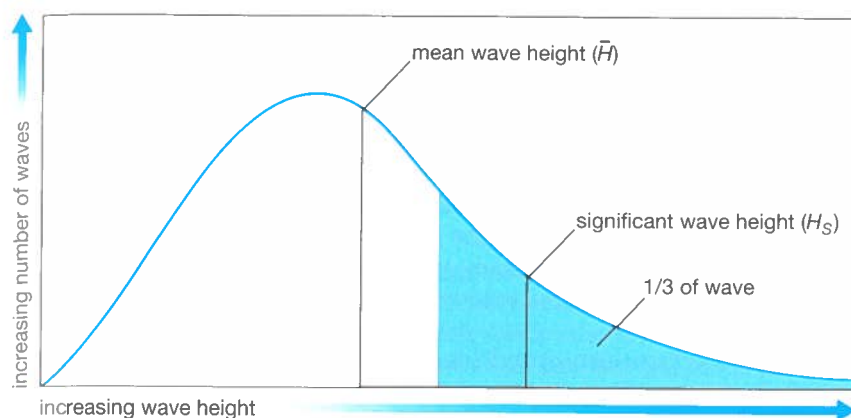


Figure 4.11 A theoretical spectrum of measured wave heights, where the average wave height of the highest 1/3 of all waves is defined as the significant wave height, or H_s .

disappears from view. The distance from the disk to the water line is noted. Then, the disk is slowly raised until it reappears and the new distance (from the disk to the water line) is recorded. The average between the two measurements is called the Secchi depth SD . Measures of Secchi depth loosely represent the maximum depth through which visible light can penetrate, with sufficient spectral quality and intensity, to support photosynthesis.

Chemical Variables

Because water is such an effective solvent, it is capable of dissolving nearly any molecule to which it is exposed. Since the world ocean has been in existence for somewhere between 3.5 and 3.8 billion years, it is no surprise that the oceans would contain such a complex mixture of chemicals, not the least of which are the salts that define our global marine waters.

Although the water is free to evaporate from the ocean and fall back to Earth as precipitation, the chemicals that are dissolved in the ocean cannot evaporate—they are “trapped” in the ocean basins and will remain there for untold epochs, unless removed by some other natural phenomenon. And so the primary challenge posed to our ocean chemists is the task of determining



FINDING HARMONY WITH THE TIDES

Periodic tidal currents can have a significant impact on the net direction and speed of oceanic currents. Tides will also have a significant impact on the overall depth of the water column, which is especially significant in shelf or coastal waters.

Luckily, the hydrodynamic effects of the tides are also monitored regularly at coastal observation stations across the globe. Typically, tide table data are cited for a particular location as a $\pm \Delta$ (in meters) from mean sea level (MSL), measured as a function of time t .

Tide tables can be used to predict the time and date of arrival for high and low tides ($\pm \Delta$ MSL) so field researchers can schedule their sampling efforts to compensate for tidal effects. They can also use the $\pm \Delta$ MSL predictions (or observations) to estimate the tidal current speed v_T as

$$v_T = \frac{(\Delta_2 - \Delta_1)}{(t_2 - t_1)}$$

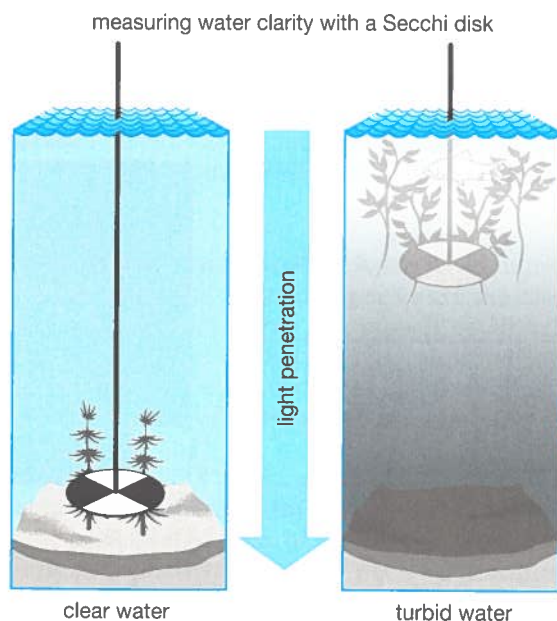


Figure 4.12 The relative turbidity of the water column will determine the depth to which sunlight can penetrate. The depth to which a lowered Secchi disk disappears from view (defined as the Secchi depth) can be used quite effectively to make comparisons of water clarity between water bodies. (Courtesy of the Minnesota Pollution Control Agency.)

both the presence and relative concentration of particular chemical species, without confusion or interference from any of the other chemicals within the same volume of seawater.

The primary constituents of seawater (other than the water itself) are the multitude of salts dissolved within. Not only do these salts contribute to the overall chemical environment of the ocean, but they also contribute to the overall density of the water. But beyond the basic measurement of the ocean's salt content (or **salinity**), there are many other chemicals present within seawater that have tremendous significance and must therefore be quantified.

Salinity Is the Fundamental Measure of Salt Content in the Ocean

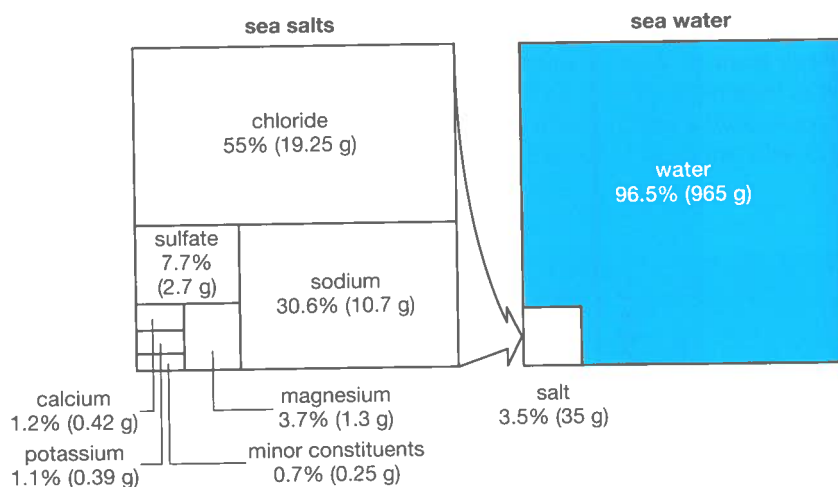
Measuring the overall salt content within a sample of seawater is not as easy as one would think. Certainly, if seawater were a simple mixture of water and a single variety of salt (like NaCl), we would simply be required to measure the concentration of either the sodium ions (Na^+) or the chloride ions (Cl^-) in solution. Unfortunately, the major constituents of seawater include several ions as well, such as sulfate (SO_4^{2-}), magnesium (Mg^{2+}), calcium (Ca^{2+}), potassium (K^+), and many others. From a chemist's point of view, this is the worst-case scenario, as this would require a different chemical test to determine the concentration of each and every ion in solution.

Fortunately, the salt ions within the world ocean are relatively **inert** and their concentrations always remain in constant proportion to each other (**Figure 4.13**). That means that no matter where a sample of seawater is taken, and no matter how salty the water is, the proportion of salt ions will always be the same: 55.5% Cl^- , 30.6% Na^+ , 7.7% SO_4^{2-} , 3.7% Mg^{2+} , and so on. Because all seawater follows this **principle of constant proportions**, ocean chemists only need to determine the concentration of one major salt ion in order to derive all the rest.

Historically, the concentration of the most dominant salt ion (Cl^-) was first determined and ultimately related to the overall salinity of the water sample. When using chemical tests for chloride concentration (or **chlorinity**), salinity ($S_{\text{‰}}$) can be determined according to the experimental relationship defined in Equation 4.15, where

$$S_{\text{‰}} = 1.80655 \text{ Cl}^- \text{‰} \quad (4.15)$$

Figure 4.13 Although salinities typically range from 0‰ in fresh water up to 38‰ in seawater, the *proportion* of salt ions always remains the same. That means a sample of 35‰ seawater will always contain 55.0% Cl^- (or 19.25 g $\text{Cl}^- \text{ kg}^{-1}$ seawater). If the salinity were 10‰ instead (10 g total salt kg^{-1} seawater), the overall amount of salt would of course be lower, but Cl^- would still comprise 55.0% of its total salt content. (Courtesy of Hannes Grobe/CC-BY-SA-2.5.)



Because there are so many different salt species present within seawater, it is more correct to consider salinity as the sum of all salt ions present within a known quantity of water. By convention, oceanographers typically use “parts per thousand” (‰) as a unit of measure for salinity, based on the mass ratio between solute (salt) and solvent (water). Thus, a 1-kg (1000-g) sample of seawater that contains a total of 35 g of dissolved salt ions has 35 g salt per 1000 g total, or 35‰.

As a practical matter, salinity is almost never measured by determining the weight of dissolved salts. Instead, most oceanographers measure the **conductivity** of sea water and ultimately relate those measurements to the overall salt content of the water sample. This practice led to the adoption of the **Practical Salinity Scale** (PSS) in 1978, when the global oceanographic community agreed to redefine the method for measuring salinity by comparing the conductivity of seawater samples against a standard salt solution of potassium chloride (KCl). Salinity measurements that are determined according to the PSS are in fact unitless; however, some oceanographers cite these salinity measures in terms of practical salinity units (PSU).

Conductivity Represents an Indirect Way to Measure Salinity Using Electricity

Because seawater contains so many dissolved salt ions, it behaves as an **electrolyte** capable of conducting electricity. Since the conductance of seawater is directly related to the amount of salt dissolved in solution, oceanographers are able to determine the salinity of seawater based on the flow of electricity through the sample. Although the best laboratory practices can usually detect salinities with 1/8000‰ precision, conductivity measurements are far superior and can detect the most subtle differences in salinity, up to 1/40,000‰! As a result, most modern methods of measuring salinity rely on conductivity meters, like those integrated in CTD recorders (see Figure 4.5).

Metabolic Nutrients Are Dissolved Molecules which are Necessary for Life in the Ocean

Beyond the inert salts that define the salinity (and density) of seawater, there are several other chemical species found in seawater, in much smaller concentrations, that have significant implications on the **biogeochemistry** of the ocean. The most important of these are the inorganic nitrogen- and phosphorus-based nutrients that are extremely **labile** and heavily utilized by living organisms to metabolize or accumulate biomass. As most biologically important nutrients are found only in trace amounts, it is always advisable to use the most sensitive tests available to determine their *in situ* concentrations.

Among the several different varieties of nitrogen-based nutrients present in the ocean, the most important chemical species are the dissolved ammonium (NH_4^+) and nitrate (NO_3^-) ions. The growth of all living organisms is largely dependent on the availability of NH_4^+ and/or NO_3^- in the ocean; therefore, any oceanographic study that investigates some aspect of biological production in the ocean would require an assessment of *in situ* NH_4^+ and NO_3^- concentrations.

With regard to phosphorus-based nutrients, the orthophosphate ion (PO_4^{3-}) is utilized almost exclusively of any other form of phosphorus. Like NH_4^+ and/or NO_3^- , orthophosphate is absolutely necessary for biological growth (making it a vital nutrient for study).

Supplemental Measures of Ocean Chemistry

Structural Nutrients Are Dissolved Molecules which Can Form Solid Mineral Deposits

Dissolved species of silicon (Si) and calcium (Ca) are absolutely necessary for the secretion of protective coatings or “shells” for some of the most important microorganisms in the ocean (which form the base of the entire oceanic food web). Dissolved calcium (as Ca^{2+}) is available as a major constituent of seawater and is used by organisms most commonly in the formation of calcium carbonate (CaCO_3) shells or skeletons. Soluble forms of silicon, such as Si(OH)_4 (silicon hydroxide), are used by certain marine microorganisms in the same proportions as NH_4^+ or NO_3^- (a testament to the importance of dissolved silica as a structural nutrient). Because of these relationships, investigations that focus primarily on the capacity for biological production will benefit from the inclusion of structural nutrients as well as the metabolic nutrients.

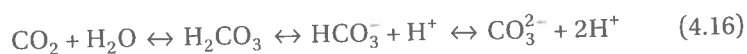
Dissolved O_2 Is One of the Most Influential Molecules that Affects the Biogeochemistry of the Ocean

Although it is perhaps easier to conceptualize how salts become dissolved in seawater, we must not forget that gases can also dissolve in seawater. The availability of dissolved oxygen (DO) gas is a critical concern for many biological investigations, particularly as it relates to the ability of seawater to sustain animal life. DO concentrations can be used to estimate gross respiration and/or photosynthetic rates (as they relate to biological processes) as well as gas exchange rates with the atmosphere (as a function of turbulent mixing, water column stability, and a variety of other physical phenomena). The availability of DO also determines the type of chemical reactions which can occur within the ocean, which in turn has a huge influence on all aspects of marine geochemistry.

Dissolved CO_2 Can Be Used to Investigate Gas Exchange and Ocean Acidification Dynamics

Carbon dioxide represents the flip side of the dissolved O_2 picture: as O_2 is consumed in the water column, it is largely replaced by CO_2 during respiration. During photosynthesis, CO_2 is removed from solution and replaced (more or less) with O_2 . By monitoring the changes in these dissolved gases, it is often possible to quantify the relative biological contributions of both the consumers and the producers in the ecosystem.

CO_2 concentrations also have a critical impact on the contributions of several different inorganic forms of carbon within the ocean, according to the carbonate equilibrium:



The most direct consequence of rising CO_2 concentrations is the liberation of H^+ , which ultimately affects the acidity (pH) of seawater, potentially resulting in ocean acidification.

pH and alkalinity: The pH of a solution refers to the molar concentration of the acidic hydronium ion, $[\text{H}^+]$, where

$$\text{pH} = -\log_{10} [\text{H}^+] \quad (4.17)$$

Although changes in pH can cause fluctuations in the carbonate equilibrium (see Equation 4.17), they can also affect the rates and outcomes of several important **hydrolysis** and **redox** reactions within the ocean. The natural

ability of seawater to negate the effects of rising $[H^+]$, sometimes called its “**alkalinity**,” can become diminished if there is too much H^+ in solution. When this happens, slight increases in $[H^+]$ can have a disproportionately strong effect on ocean pH, exacerbating the effects of ocean acidification.

Biological Variables

Of all the oceanographic variables to consider, those of biological significance are perhaps the most difficult to quantify. Living organisms are adaptive by nature and exhibit behaviors that can complicate our efforts to collect those data we find most pertinent. Biology is anything but passive, and rarely do we have the luxury of performing a particular measurement that doesn't affect the organism and induce a change in the very parameter we were trying to nail down. That being said, it is possible to reduce the most basic biological investigations down to just a few of the essentials.

Regardless of whether you are interested in the biology of the oceans from the zoological perspective of a single species or the ecological perspective of entire communities of organisms, the essential measures are consistent. In very general terms, we are most commonly interested in three basic biological questions: 1) who is there, 2) how many are there, and 3) what are the indicators of organism or ecosystem function? The first two are easy—it's that last one that poses the most significant challenges.

Species Identification Is the Only Way to Describe the Diversity of Life in the Ocean

In the context of natural selection and evolution, if more than one species exist in a given habitat, we must assume that they possess enough differences in their ecological functionality to have evolved into different species in the first place. Why evolve into different species if you utilize all the same resources, occupy all the same habitats, and serve all the same functions? Therefore, the simultaneous presence of multiple species must be indicative of their biological differences. By simply identifying what species are present, we are much better equipped to investigate what those differences might be.

The simplest method of species identification is to assess **species richness**: a simple count of all the different species collected (or found) in the area of interest. Although it is always best to identify organisms to **genus** and **species**, such efforts can be time consuming and require a certain degree of expertise in **taxonomy**. A collection of organisms can still be categorized into different “species,” even if the true taxonomic names are unknown. For example, you may not know the scientific names of an Atlantic flounder and a striped anchovy, but you could very easily determine that they are indeed two very different species of fish.

Species Abundance and Biomass Measures Are Used to Quantify the Amount of Biologic Matter

When assessing species abundance, there are two very basic strategies: either we can choose to perform counts of whole individuals or we can measure their **biomass** (and if the resources are available to do so, we might measure both). There is no universal rule as to when each strategy should be employed; however, the chosen expression of biological abundance should adhere to the context of the study. For instance, the results of a study focused on the disease rates of brown pelicans would be far more meaningful if we cited our results in terms of the number of diseased individuals per 1000 pelicans. However, if that same study sought to associate disease rates as a function of pelican body weight, we would be wise to collect the biomass data from those diseased and healthy pelicans.

When it is necessary to take counts of individual organisms, those data are most meaningful when the counts are related to the area (or volume) from which they were taken. For benthic organisms, areal abundances are the norm (number of individuals per square meter). For organisms surveyed within the water column, such counts are typically reported as volumetric abundances (number of individuals per square meter, or number of individuals per litre).

When it is necessary to gather biomass data, it is also important to consider whether such information will be collected from a **wet weight** versus **dry weight** perspective. The determination of wet weight is by far the simplest: the organism is simply blotted dry (or the surrounding water is eliminated by vacuum filtration) and weighed. Another advantage of assessing wet weight is that the organism can survive the procedure (if the investigator is both quick and gentle about it). Unfortunately, many aquatic organisms can retain a significant amount of fluid within their bodies, and the amount of fluid can be highly variable among different individuals of the same species; therefore, wet weights can be a misleading measure of the specimen's true biomass. A more accurate method of measuring true biomass is to first **desiccate** the specimen (in a 70°C drying oven) and then measure the organism's dry weight biomass.

Occasionally, it may be preferable to determine the dry organic weight of a specimen (as a measure of the specimen's total organic content). This is done by first determining the specimen's dry weight biomass, and then placing the specimen in a 500°C furnace to thoroughly combust all organic carbon to CO₂. The remaining ash is allowed to cool and then weighed, so that the difference between the pre- and post-combustion mass can be determined as the specimen's dry organic weight (or **ash-free dry weight**).

Species Biometrics Represent Those Measures that Suggest a Specific Biological Function

For most organismal studies, it is not enough to simply account for species richness and species abundance. For those investigations that seek to measure an organism's function (either as an individual or within the larger context of its role in the ecosystem), it is necessary that we decide on the appropriate **biometric** that represents the best anatomical or biochemical indicator(s) of that function. A fundamental assumption in biology is that all organisms exhibit adaptations that define their overall fitness for survival. If we seek to explore their unique function (that is, their unique fitness for competition and survival), we should be able to find some measurable feature that is indicative of their capacity for success within the ecosystem.

It is impossible to offer blanket guidance on an "appropriate" biometric for all investigations. The broad diversity of marine organisms, from bacteria to blue whales, compels us to consider a much larger panoply of biometrics with which to describe organism function. As a general rule, anatomical biometrics typically include measures of body or cell size and mass. Although these measures are largely regarded as "must-haves" for any investigation, they are also very general and may not be truly indicative of the specific function being investigated.

From a molecular perspective, it is even more difficult to suggest a universal biometric. However, one of the most fundamental molecular measurements, performed routinely in most oceanographic studies, is the assessment of *in situ* concentrations of **chlorophyll *a*** (chl *a*). Since chl *a* is the primary biochemical pigment that enables carbon fixation, it is found in every single organism capable of performing photosynthesis. Because of that universality, measures of chl *a* concentration are directly indicative of

the photosynthetic biomass and the overall potential to add new biomass to the ecosystem, at the very base of the aquatic food web.

Supplemental Measures of Ocean Biology

With the immense diversity of life present in the world's oceans, it is impossible to issue a "one-size-fits-all" declaration of the important biological variables to consider. However, in addition to the essential measurements suggested previously in this section, the following can be used to build on those foundations, often as a simple augmentation to the field efforts in which we are already engaged.

Accessory Pigments Can Be Used To Identify Different Species of Plants and Algae

In addition to chl *a*, photosynthetic organisms also possess a tremendously diverse assortment of other photosynthetic and photoprotective pigments (such as chlorophyll *b*, beta-carotene, zeaxanthin, phycocyanin, etc.). Since many taxonomic groups can be identified by comparing the type (and cellular abundance) of these different pigments, it is possible to use pigment analysis to identify the taxonomic groups present and their relative abundances. Since photosynthetic organisms also possess the ability to alter their internal pigment ratios in response to the available light, detailed analyses of accessory pigments can also be used to document the photosynthetic status of primary producers.

Organic C-N-P Ratios Define the Nutritional Status of the Living Biomass

Although chl *a* and accessory pigments are only found in photosynthetic organisms, all the world's life-forms are made from the elements of carbon, nitrogen, and phosphorus. Organisms growing at peak nutritional efficiency will typically accumulate organic material with a C-N-P ratio of 106:16:1, also known as the **Redfield ratio**. When organisms are surviving in nutrient-stressed conditions, they will accumulate C-N-P in very different ratios. By assessing the organic C-N-P content of a particular organism (or group of organisms), it is possible to determine the nutritional conditions under which they have been living—information that can have rather critical implications for the transfer of those nutrients (and food energy) to higher-level consumers in the food web.

Age Structure and Size Class Measures Establish the Demographics within a Specific Population

Many of the ocean's creatures begin their lives as microscopic larvae and grow, sometimes to remarkable size, according to very regular developmental stages. Therefore, it is often possible to determine the breeding histories of individual species based on the relative abundance of specimens collected in each size class (Figure 4.14). For many species, the time needed to develop into each particular life stage is well-known, so the size classes can be easily correlated with the age structure of the population. Such information can be particularly valuable for stock assessments of ecologically and commercially important species.

Geological Variables

The pertinent spatial and temporal scales in geological oceanography can run the gamut and easily stretch to the extremes. After all, many geological investigations deal with entire ocean basins or entire continental margins, and for many geologic processes, we measure the passage of time in millions

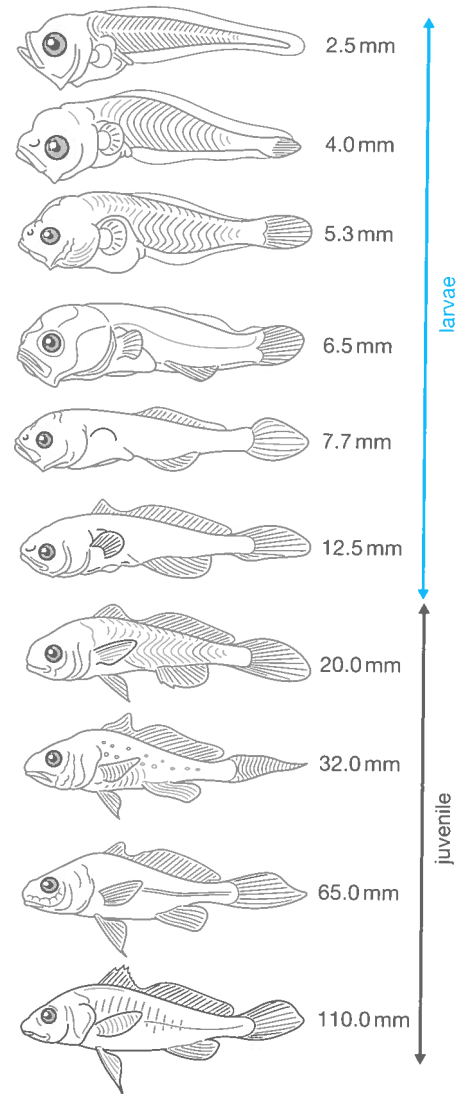


Figure 4.14 Developmental stages and size classes of the Atlantic croaker, *Micropogonias undulatus*. Since many marine species possess multiple larval and juvenile stages of growth that are highly time dependent, this information can be used to back-calculate breeding periods and to estimate maturation rates (age structure) of a given population, based on life-stage collections performed in the field. (Adapted from, Johnson GD [1978] U.S. Department of the Interior).

(or perhaps billions) of years! Although such geological studies may incorporate field data collected across broad geographic regions, it is perhaps more useful for us to limit our perspective to “smaller-scale” geologic investigations that may more easily be accomplished by one (or few) researchers in a localized study.

For local- or regional-scale investigations in ocean geology, perhaps the most popular areas of interest involve the influence of the coastal margins and the ocean bottom. As these quite literally represent the physical boundaries of the ocean basins, their influence is decidedly geologic in nature, largely as a function of their **geomorphology**, depth, and sedimentary or depositional character.

Geomorphology and Bathymetry Describe the Complex Geometry at the Ocean Boundaries

Because it is so much more convenient for us to perform oceanographic research along (or within a short distance from) the coastal margin, the geomorphology of the coast plays a critical role in virtually every ocean dynamic we can consider. Defining the physical boundary of the ocean basin in the horizontal dimension is foundational to our frame of reference for any oceanographic study; without it, we cannot define what might be considered “along-shelf” or “cross-shelf” processes. As simple as it sounds, the orientation of the coastline defines our onshore and offshore directions and helps us establish our fixed Eulerian frame of reference.

Just as the geomorphology helps us to establish our horizontal boundaries, a high-precision survey of ocean depth (or **bathymetry**) establishes the vertical limits of the water column and has profound significance on virtually every aspect of ocean science we might choose to study (particularly when it comes to phenomena that are influenced by the ocean floor). Thus, it is always advisable to acquire the most recent, high-precision bathymetry data for the region of your oceanographic study.

Suspended Sediments Provide the Critical Link between Land and Sea

Beyond the physical boundaries of the coast and the ocean bottom, one of the most important geologic aspects in an area is related to the deposition of sediments. Sediments that are suspended in the water column can significantly affect the penetration of sunlight, alter the chemical environment due to the slow dissolution of suspended particles, or influence the survival and success of different organisms that cannot easily cope with the suspended sediments. As most of these suspended particles are more dense than the surrounding water, they will eventually settle to the ocean floor. Sediment deposition can, over time, significantly affect the bathymetry of a region. The accumulation of sediment particles on the ocean floor can also have dramatic influences on the chemical environment as the ocean-sediment interface, or perhaps even cause the burial of benthic organisms. In fact, the simple rate of sediment deposition can provide valuable insight into the depositional environment of a region.

Typically, the deposition of sediment is measured by deploying subsurface sediment traps (**Figure 4.15**). Collection vessels, fitted with special funnels to collect the sediments settling into the collection chamber (but exclude stray currents or browsing organisms), are deployed at different depths for a fixed period of time. Since the amount of sediment collected is proportional to the size of the sediment trap’s mouth (and to the duration of collection), it is a relatively simple matter to determine depositional rates.

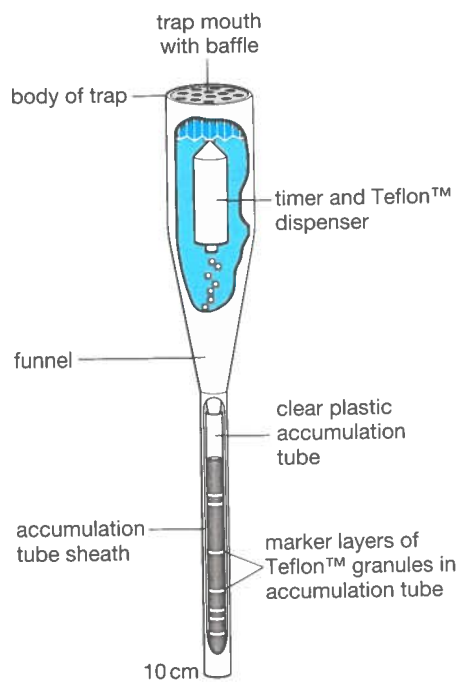


Figure 4.15 A schematic drawing of a time-marking sediment trap. The baffles in the trap mouth prevent turbulent mixing of the sediments and exclude unwanted organisms from the accumulated material. In this design, an automatic timer drops Teflon™ granules into the sediment at defined time intervals to mark the time of deployment.

Sediments can also be studied after they have been deposited on the ocean floor. By using any number of coring devices (**Figure 4.16**), the many layers of deposited sediments can be recovered while maintaining the integrity of the sediment layers. This method is particularly effective in studying past depositional environments, as the layers of sediment within the core can be analyzed to piece together the depositional history of the region, as recorded in the various sediment layers.

Supplemental Measures of Ocean Geology

In support of virtually every type of geologic collection method, measurements of the physical dimensions of the sediment grains, as well as their chemical content, can add tremendous insight to the depositional environment as recorded by sediment traps or cores.

Sediment Grain Characteristics Provide Forensic Evidence of Historical Events

Sediment grains collected within sediment traps or cores can be chemically analyzed for virtually any atomic element or molecule of interest. Depending on the nature of the study, such analyses can be used in the forensic analysis of the mineral source of the deposited material, to track changes in the chemical constituency of deposited sediments over time, or to investigate the organic content of the sediments (as evidence of biological activity).

Microscopic analysis of grain shape can help geologists establish the relative age (or intensity of erosion) of the deposited sediments. Depending on the mineralogy and “hardness” of the sediment grains, angular and irregularly shaped grains are indicative of relatively young sediments, whereas those that have been eroded more intensely (or for a longer period of time) tend to be more regular in shape, with far more rounded edges (**Figure 4.17**).

Grain size can be determined even more easily by simply passing a known mass of sediment through a series of sediment screens (or sieves) with progressively smaller mesh sizes (**Figure 4.18**). In this way, sediment grains can be grouped into different size classes based on how much of the total sediment sample has been retained in each of the different sieves.

Other Variables

The major oceanographic variables have been discussed, but there are a few meteorological parameters that are routinely measured owing to their profound influence on various ocean processes. Although most of these meteorological variables primarily affect the physical aspects of ocean dynamics, their significance in the chemical and biological realms cannot be overlooked.

Winds Provide a Strong Influence on Flow Fields and Surface Currents

One of the most significant drivers of ocean surface waves are the winds. Physical variables such as significant wave height and current velocity are largely determined by wind speed, and the direction of the winds (relative to any impediments to flow, such as islands or coastlines) can have dramatic impacts on the dynamics of water transport. Fortunately, measures of wind speed and direction are almost universally collected from coastal and ocean observing stations across the globe.

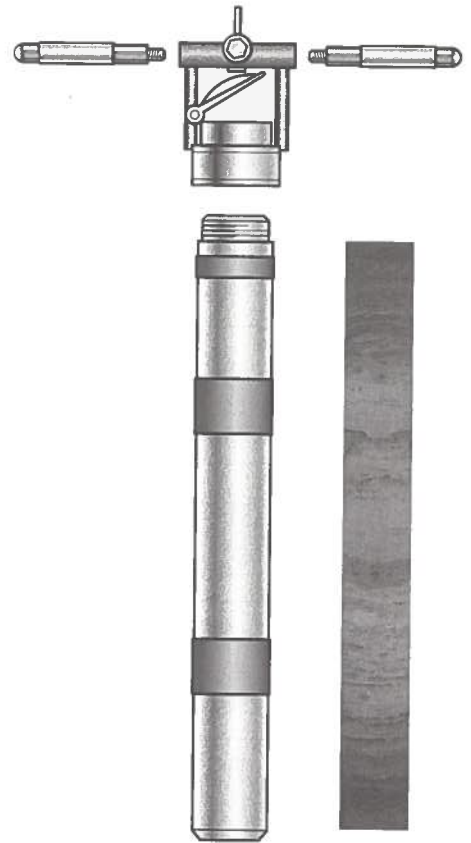


Figure 4.16 A schematic drawing of a push corer (*left*), a device designed to recover sediment cores without disturbing the delicate layering of the deposited sediments. Each sediment layer in the sediment core (*right*) can then be analyzed separately to determine its unique depositional characteristics (for example, grain size, shape, mineralogy) as well as to determine when each layer was deposited.

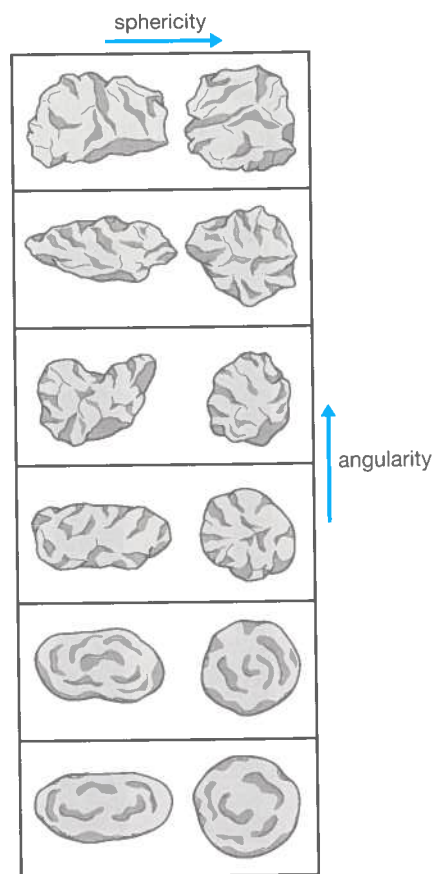


Figure 4.17 Sketches of microscopic analysis of both the angularity and sphericity of sediment grains within each layer of a sediment core can yield a tremendous amount of information about the age of the sediments, as well as the erosional intensity prior to deposition. (Adapted from Powers, MC [1953] *Journal of Sedimentary Petrology*, 23(2):118.)

Figure 4.18 A series of sieves with progressively smaller mesh sizes, commonly used to sort sediments according to their size class.

Barometric Pressure Can Affect the Hydrostatic Pressure of Ocean Water Beneath It

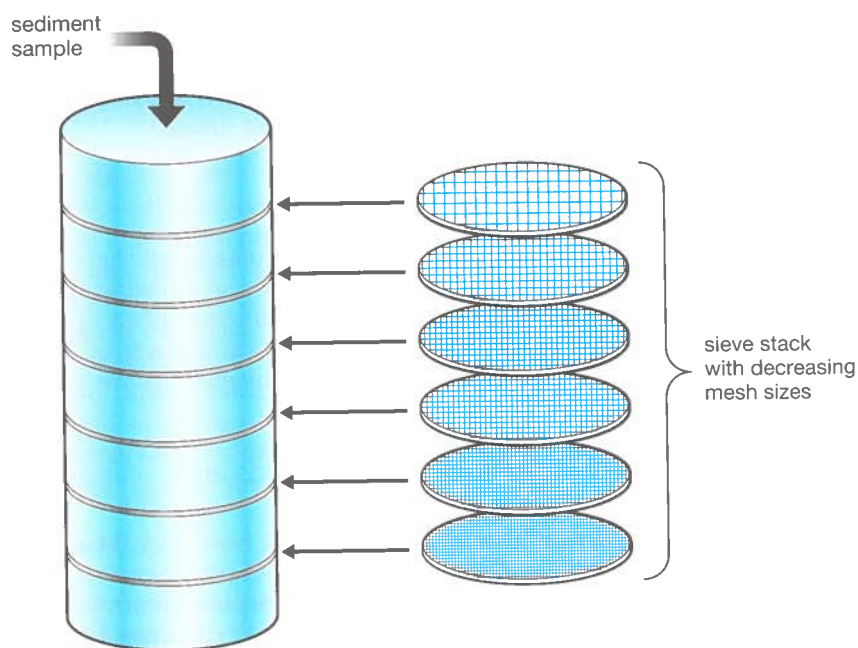
Like wind speed and direction, barometric pressures are routinely measured at weather stations located at sea and along the coast in virtually every location on the globe. If high-precision calculations of hydrostatic pressure are required, it will be necessary to use barometric pressure data to correct for any departures from the “standard atmospheric pressure” of 101.325 kPa, which is assumed to be constant to establish a hydrostatic pressure of $P = 0$ for the Equation of State of Seawater (IES 80; see Technical Box 4.1)

Air Temperature Can Provide a Delayed Influence on Water Temperature

Like wind velocity and barometric pressure, air temperatures are also routine measures at coastal and oceanic weather stations. Because seawater is slow to gain and lose heat, the effects of air temperature on ocean dynamics can largely be ignored if water temperatures are known. However, changes in air temperature can have a significant effect at the air-sea interface and lead to heating or cooling of the uppermost layers of the ocean (which can lead to increased mixing and overturn). Certainly, extreme seasonal variations (or transient temperature fronts) can lead to significant heating or cooling of surface waters.

Precipitation Will Affect Both the Temperature and Salinity of Surface Water in the Ocean

Although the quantity (and type) of precipitation is routinely monitored at land-based weather stations, it is seldom monitored at offshore stations. Depending on the nature of the study, precipitation can have a significant impact on surface salinities and may even increase the surface concentrations of certain nutrients and salts that are scoured from the atmosphere and delivered to the ocean surface during precipitation events.



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