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Barry A. Klinger , Thomas W. N. Haine
Excerpt
More Information

1 Physical Oceanography: Methods and Dynamical Framework

Physical oceanography extends well beyond the study of the general ocean circulation. Physical oceanographers investigate tides, wind-driven surface waves, air/sea interaction, sound in the ocean, light in the ocean, and physical processes at the sea bed, on beaches and on man-made structures like breakwaters and ships. Physical oceanography is also key to chemical and biological oceanography, and marine geophysics. Through its coupled interaction with the atmosphere, cryosphere, biosphere, and land, the physical state of the ocean also plays a critical part in climate dynamics and Earth system science.

Nevertheless, study of the ocean general circulation is central to physical oceanography and is playing a central role in the development of the field. This development involves an interplay between physical observations of the ocean, theories of the fluid dynamics, and numerical models of the circulation. The focus in this book is on the time-averaged, global ocean currents. In order to approach this topic we must set the stage by reviewing some preliminary ideas about physical oceanography. We are selective in the coverage, only discussing topics that recur later.

For comprehensive discussions of empirical methods in physical oceanography, consult the books by Emery and Thompson (2001) or Wunsch (2015). For more details on dynamical oceanography consult Vallis (2006), Marshall and Plumb (2008), or Pedlosky (1996), among others. For reflections on the field's history, Jochem and Murtugudde (2006) and Mills (2011) are good starting points .

1.1 Observations

1.1.1 What Is the General Circulation?

The **ocean general circulation** is a description of the three-dimensional velocity field of the ocean, $\mathbf{u}$. It is difficult to define the concept more exactly, or to precisely distinguish it from the rest of physical oceanography. Nevertheless, it essentially means *the slowly changing, large-scale velocity field*. By "slowly changing" and "large-scale" we usually mean timescales of seasons and longer, and length scales ranging from about 100 to 10,000 kilometers.

In discussing the general circulation, we include other key physical properties such as temperature, salinity, density, sea-surface height, and pressure. These properties are intimately related to the dynamics and thermodynamics which determine the velocity field itself. Physical oceanographers study how all these physical properties vary from place to place, and analyze physical principles which underlie the ocean behavior of different regions of the ocean. Application of these principles leads us to try to understand the ocean circulation on global scales.

Most of this book discusses an abstraction known as the **steady** circulation. We know that the circulation is not steady and actually varies on a wide range of timescales. For instance, observations of the eastward component of velocity in the equatorial Pacific show variations from year to year, (Figure 1.1a), month to month (Figure 1.1b), day to day (Figure 1.1c), and even within an hour (Figure 1.1d). Despite this temporal variability, Figure 1.1 shows that the current (the equatorial undercurrent) at that particular location is most often eastward with strength of around 0.6 to 0.9 m/s (Subsection 6.1.1).

How does this current compare to flow at other locations? The most intuitive way to compare is simply to take a **time average** over a long time. How long that time should be is not well defined, because, as we discuss in Subsection 1.2.1, ocean characteristics vary from seconds to geological timescales. As with the short time variations shown in Figure 1.1, sporadic measurements have indicated that many physical ocean features are stable over the last century or so. In that case, averaging over any time interval from a few years to a few centuries will give a similar answer, so the precise averaging interval is not so important. However, the more precisely we try to define the circulation, the more sensitive our answer is to the averaging interval.

Such a long-term average often produces what is known as a **climatology**. It is analogous to the original meaning of **climate** as a description of long-term averages of atmospheric characteristics. The atmospheric meaning of climate often includes physical ocean characteristics because these play an intimate role in determining atmospheric conditions.

Sometimes the long-term average circulation is considered a synonym of the general circulation, but a broader definition usually includes some measure of time variability. Like the atmosphere, the ocean has many features that strongly depend on the time of the year, so a climatology is often defined as the average (over many years) of the evolution of a variable over the course of the year. For other timescales, it is the *statistics* of the variability rather than the detailed evolution over time that we usually consider part of the general circulation. For instance, we think of the mean value and the standard deviation of the time series of the equatorial undercurrent data (see the histogram in Figure 1.1).

Similar issues arise when we consider spatial variations in the general circulation. Ocean parameters such as current speed vary on all length scales from millimeters to the width of the basin (Subsection 1.2.1). Taking averages over multiple years tends to eliminate features with horizontal scales less than a few hundred kilometers. Larger scales are affected by smaller-scale steady structures in some regions and by the statistics of time-varying small-scale structures in many areas (Chapter 7).

Like all physical science, physical oceanography is an empirical discipline, but one in which performing controlled experiments is generally impossible because of the great scale

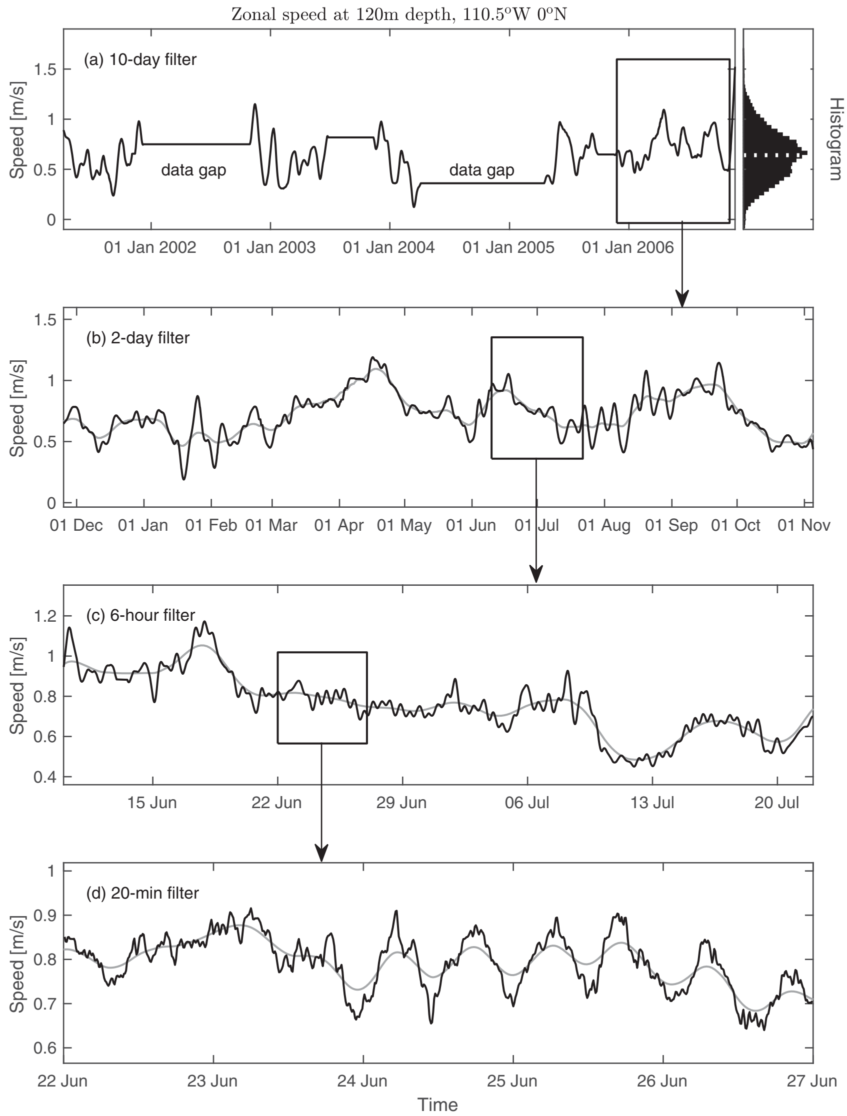

Figure 1.1 Time series of zonal current at 120 m depth in the equatorial Pacific ocean. Several different periods are shown, ranging from days (bottom) to years (top). At top right is a histogram of the speed data with the mean shown by the dotted line. Data are from the OceanSITES Tropical Atmosphere Ocean/Triangle Trans-Ocean Buoy Network moored buoy array (Appendix A).

of the ocean and ethical considerations. Also, we do not have additional oceans to tinker with! These limitations are shared with other fields in Earth and space sciences. Our knowledge comes from careful observations of the natural system evolving in its own way. However, laboratory experiments in rotating tanks of water provide analogues to ocean phenomena upon which we can conduct experiments.

Table 1.1 Comparison of sensor technologies for observing physical properties of the ocean.

	Property	Unit	Accuracy	Sensor
	In-situ properties:			
p	pressure	Pa, db	0.001–0.1 db	strain-gauge sensor
T	temperature	°C	0.001 °C	thermistor
S	salinity	(none)	0.005	conductivity sensor
ρ	density	kg m^{-3}	0.01 kg m^{-3}	(from T, S, p)
$\mathbf{u}$	velocity	m s^{-1}	O(0.01) ms^{-1}	current meter
χ	tracer	Mol kg^{-3}	varies	sensor or lab. analysis
	Remote-sensed sea-surface properties:			
η	height	m	3 cm	radar altimeter
T	temperature	°C	0.7°C	IR & microwave radiometry
S	salinity	g/kg	0.1–0.2	microwave radiometry
	Navigational position:			
H	water depth	m	O(1) m	Echo sounder, satellite
p	pressure	Pa, db	0.001–0.1 db	strain-gauge sensor
ϕ	longitude	degrees	5 m	Global Positioning System
θ	latitude	degrees	5 m	Global Positioning System

1.1.2 Sensors to Observe the Ocean

Observations are a key part of physical oceanography. Here we summarize how to make physical observations of the ocean. First, we discuss sensor technologies, then we discuss instrument platforms. Table 1.1 summarizes this information. The notation we introduce here is used throughout the book, and is also summarized in the tables in Appendix C, which contains a useful summary of numbers relevant to physical oceanography. Appendix A is a reference for the datasets we show throughout the book.

Sensors to observe the ocean are diverse and constantly being improved. Our brief discussion splits sensors into two categories: in-situ sensors, which are physically in contact with the water, and *remote* sensors, which observe the ocean from above or by acoustical methods.

In-Situ Sensors

In-situ instruments measure physical properties of seawater by direct immersion in the sea. The most important properties measured this way are:

Pressure (p, see Table 1.1) is the force per unit area exerted by seawater at a particular place in the ocean and is a key quantity in understanding currents in a fluid. The force is measured perpendicular to a specific flat surface, but it does not depend on the orientation of that surface and is thus a scalar. (More generally, pressure arises from the normal components of the stress tensor in continuum mechanics. Detailed knowledge of tensors is not required for this book, although they are used in Chapter 7; Griffies, 2004 provides details.) We can think of pressure as the force of molecules banging against each other. Pressure does not have a direction because pushing in one direction on a volume of water causes the force to be transmitted in all directions. This is illustrated by a vertical tube of water, in which the pressure is due to the downward force of gravity but the water exerts a pressure force on the inside of the tube, perhaps even breaking the tube if the force is strong enough.

Various units of pressure are commonly used. The SI unit is $\mathrm{N\,m^{-2}}$ or Pascals (Pa; $1\ \mathrm{N\,m^{-2}} = 1$ Pa), but oceanographers often use decibars (db, dbar; 1 tenth of a bar, where 1 bar = $10^5\ \mathrm{N\,m^{-2}}$). A pressure sensor is typically made with a silicon strain gauge or an oscillating quartz crystal. Modern instruments have accuracies of around 0.02–0.1% of their full design range of pressures, and full-ocean depth coverage is possible. Long-term sensor stability is about 0.02–0.1% of full-range pressure per year. Although this is a small fraction, it corresponds to a water column uncertainty of a few meters each year, which is a large dynamical signal (we explain why in Subsection 2.2.2).

The value of the pressure is given accurately by the **hydrostatic balance**, which equates the seawater pressure to the weight of water overhead (see Subsection 2.1.2). The hydrostatic balance is an excellent approximation for the large scales we focus on in this book (Section 1.1), though it breaks down for phenomena characterized by length scales of less than a few kilometers and timescales of less than a day. Observational oceanographers usually use pressure – which is easily measured – as a surrogate for depth – which is not. Because density is nearly constant in the ocean (Subsection 1.2.3), converting pressure to depth is straightforward: roughly speaking, 1 db corresponds to a vertical water column of 1 meter. Sea-level atmospheric pressure is typically close to one "standard atmosphere," which is 101,325 $\mathrm{N\,m^{-2}}$, so 1 db$\approx$1 m$\approx$10% of a standard atmosphere. In other words, the weight of 10 m of seawater causes a pressure equal to about 1 atmosphere (a fact familiar to all scuba divers). Ignoring the small variations in density, and some other subtleties concerning hydrostatic pressure, can cause errors of up to a few meters in equating depth and pressure. These errors are small compared to the depths of most features we will discuss, which are typically hundreds of meters or more.

In-situ temperature (T, Table 1.1) is measured using a thermometer in contact with the water. Typically, a thermistor (a thermal resistor) is used in oceanography. Mercury-in-glass thermometers were widely used before the era of micro-electronics began in the 1960s, but are now mainly museum pieces.

Moving a seawater sample deeper without exchanging any heat with the surroundings (an **adiabatic** movement) increases its pressure, and hence slightly compresses the sample. This squeezing does work on the sample and therefore increases its temperature. Similarly, lifting a sample to the surface decreases the pressure, causing the sample to expand and therefore cool. The atmosphere behaves similarly but with greater temperature change because air is much more compressible. **Potential temperature** (θ) is the temperature a sample would have if it were adiabatically moved to a **reference pressure**, typically zero pressure, taken to be at the sea surface. Potential temperature allows easy comparison of temperatures for water parcels at different pressures. Most often we want to know how the temperatures would differ if the parcels were brought to the same pressure as each other, so in this book we use potential, rather than in situ, temperature. Potential temperature θ is calculated from T using an empirically derived formula that mainly depends on T and p, and is a few tenths of a degree Celsius cooler for full ocean depth, as seen in Figure 1.6a. Where the context removes ambiguity, we use T to mean potential temperature rather than θ. Potential temperature is now being superceded by **conservative temperature** Θ (Subsection 1.2.3), which is essentially similar for our purposes.

Salinity (S, see Table 1.1 and Subsection 1.2.3) measures the concentration of dissolved salts in seawater. Salinity varies in the ocean due to concentration (from evaporation) and dilution (from precipitation and river runoff), but the abundance of different salts (sodium chloride, potassium chloride, etc.) relative to each other is nearly constant throughout the ocean. The most recent definition of salinity, called the **absolute salinity**, S_A, is the mass of dissolved salts per mass of seawater (Equation 1.7). Thus, $S_A = 35$ g/kg implies a concentration of 35 g salts in 1 kg of seawater. Much of the oceanographic literature uses the 1978 practical salinity scale (pss-78), which is based on seawater electrical conductivity and measured in practical salinity units (psu) or parts per thousand ("ppt," "per mille," or "‰"). Here we ignore the numerically small differences between absolute salinity and practical salinity.

Salinity is measured by oceanographers using an immersed conductivity cell. A thermistor and a pressure gauge for simultaneous temperature and pressure measurement are also often needed. Such observations of temperature and salinity at different pressures are called **hydrographic** observations (literally, mapping of water properties). Laboratory calibration of a sensor using standard seawater solutions is required for best results. A key challenge for long-term deployment of conductivity cells is to avoid fouling of the cell by biological material. The long-term drift of modern salinity sensors is 0.05 g/kg or better over several years, however.

Seawater **density** (ρ, see Table 1.1, Subsection 1.2.3) is the mass of a seawater sample divided by its volume, and is a critical variable in physical oceanography. Density is not measured directly at sea. Instead, density is inferred from the temperature, salinity, and pressure of a seawater sample using an empirical formula called the **equation of state**. The formula is estimated experimentally using laboratory samples of seawater and highly accurate analytical instruments. The accuracy of density estimates using in-situ temperature, salinity, and pressure measurements is around 0.01 kg m^{-3}, the main source of error being uncertainty in the equation of state itself.

Velocity (**u**, see Table 1.1) is the current vector of the fluid flow, and is measured directly with a **current meter**. The many types of current meter can be broadly categorized as mechanical, acoustic, or electromagnetic devices. Mechanical current meters often consist of a rotor whose spin is proportional to current speed, and a vane to turn the meter to the current direction which is measured by a compass. Mechanical current meters have been used for many decades (see Figure 1.2) and are often deployed on moorings (Subsection 1.1.3) which are designed to hold the current meter at a fixed position in the water. A modern mechanical current meter (for example, the Aanderaa RCM-8) can measure currents from 0.02 m/s (below which the rotor stalls) to about 3 m/s, with an accuracy of the greater of 0.01 ms^{-1} and 4% of the measured current.

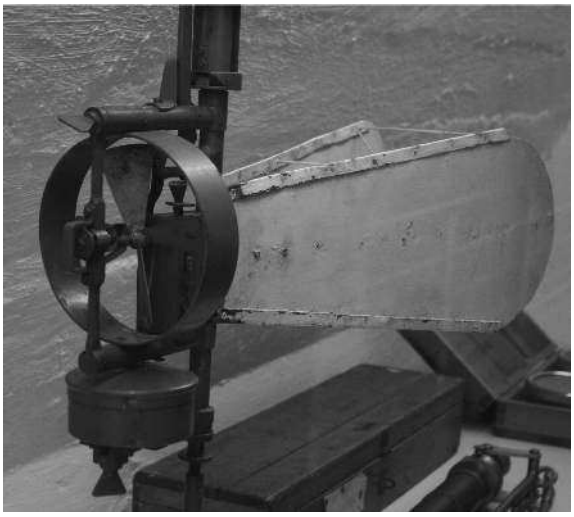

Figure 1.2 A mechanical current meter constructed by Vagn Walfrid Ekman in the early twentieth century. The essential design, such as the rotor and vane, is still in use today. From the Frammuseet, Oslo.

Many modern current velocity measurements are made with **Acoustic Doppler Current Profilers** (ADCPs), which exploit the Doppler effect on high-frequency sound waves reflecting off plankton and other suspended material drifting in the current. The shift in frequency of the reflected sound yields the speed of the flow. The ADCP uses a range of return times from sound radiated in different directions to estimate the three-dimensional velocity vector over a range of depths. Ranges of several hundred meters, resolution of a few meters, and accuracies of a few cm^{-1} are common. Figure 1.1 shows data from an ADCP attached to a mooring (Subsection 1.1.3) which was recovered and redeployed several times over the decade.

Some current meters are electromagnetic and work on Faraday's principle of induction for a moving conductor (seawater) in the Earth's magnetic field. This type of instrument is well adapted to turbulent near-shore environments, in the presence of surface waves, for example, and regimes with very low or very high currents. They are not often used for current measurements in deep water environments, however.

Finally, we refer to a whole class of in-situ methods as **serendipitous current measurements**. These include mariner's reports, tracking the location of drifting objects, either at the surface or at depth, and interpreting the spread of natural, or deliberately released, chemical **tracers**, such as dissolved oxygen gas. The surface ocean circulation was described in a basic way using these methods long before systematic scientific instruments were deployed. Indeed, chemical tracers are still used to give information about deep currents. Some kind of "clock" is needed so that changes in tracer concentration from place to place, or time to time, can be associated with the ocean current. For example, radioactive decay of carbon-14 shows that the deep Atlantic is much "younger" than the deep North Pacific. Man-made compounds like chlorofluorocarbons (CFCs) with known transient atmospheric time histories are another example, and have been used since the 1980s to trace deep and abyssal flows. A large database of such tracer measurements now exists for the global ocean (see Figure 1.4). Interpreting the tracer data in an unambiguous way is not always easy, but they provide information on ocean circulation that complements that from hydrography and current meters.

Remote Sensors

Remote sensors operate by probing the ocean with some sort of wave. Acoustic waves are a good example because the ocean is transparent to sound. An ADCP (see above in this subsection) exploits this fact to measure currents at a distance from the instrument, and an echo sounder measures water depth this way. Acoustic networks have also been deployed to monitor average ocean temperature over basin and even antipodal scales. These networks have not yet been operated in an ongoing mode, however. See the book by Munk et al. (1995) for more details (and Wunsch, 2015; Dushaw and Menemenlis, 2014); we do not discuss acoustic methods further here.

The ocean is opaque to electromagnetic waves, which means that remote sensors observing the ocean from satellites (or aircraft) are generally limited to measuring surface properties. Nevertheless, observing the ocean from space using satellites has revolutionized physical oceanography in the last 30 or 40 years. Satellites provide much more comprehensive coverage in space and time than in-situ instruments. The most important satellite *remotely sensed* properties are as follows:

Sea-surface temperature (SST) has been measured over the globe from space since 1970. The technique involves observations of infrared radiation from the Earth's surface in multiple wavebands. There are some drawbacks, however. First, the infrared radiation observed by the satellite is emitted from a thin surface layer 10–20 $\times 10^{-6}$ m thick. This layer is colder than the bulk liquid by O(0.1)°C, but the difference varies with environmental conditions and can be hard to estimate (see Subsection 4.1.1). More importantly, clouds block infrared radiation from the sea surface, so many SST measurements are obscured. Some satellites avoid this problem by looking through cloud with microwave radiation rather than infrared, but the spatial resolution and data accuracy are not as good.

Nowadays, data centers provide daily global SST maps based on merged observations from multiple satellites. The coverage is at a resolution of up to about 5 km with an accuracy of about 0.7°C. These data are invaluable to oceanographers and climate scientists, and SST is the best-measured parameter in marine science. An example of a satellite SST map is in Figure 4.1.

Recently, remote sensing of **sea-surface salinity** (SSS) has become practical through the measurement of microwaves emitted by the sea surface combined with SST and surface roughness measurements. One of the first major missions, Aquarius/SAC-D, was operated from 2011 to 2015 and achieved 0.1–0.2 g/kg accuracy with a spatial resolution of 100–300 km and a repeat cycle of 7–30 days. Aquarius collected as much salinity data as the entire historical data base many times over and is becoming an important

Schematic of Satellite Altimetry

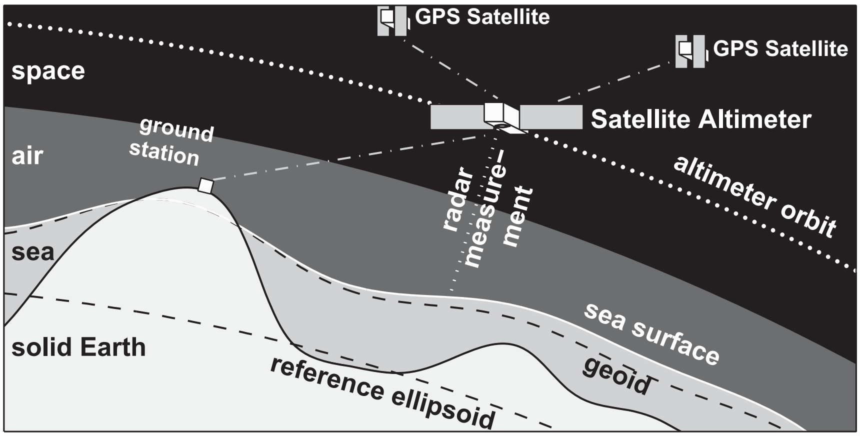

Figure 1.3 Schematic of SSH measurement with TOPEX/Poseidon satellite altimeter. Image from NASA Jet Propulsion Laboratory.

new data source to oceanographers. An example of a sea-surface salinity map is in Figure 4.4.

Sea-surface height (SSH, see Table 1.1 and Figure 1.3), sometimes called ocean surface topography, is measured by an altimeter, either flying on an aircraft or a satellite. The instrument uses radar to measure the distance between itself and the sea surface. Given accurate information about the instrument position, the height of the sea-surface can be estimated to within a few cm. Every ten days, altimeters TOPEX/Poseidon (1992–2005) and Jason 1 and 2 (2001 to present) have been covering most of the global SSH field with 7 km resolution along tracks that are 315 km apart at the equator.

Sea-surface height is a critical quantity to physical oceanographers because it is directly connected to the surface ocean current via the simple **geostrophic relation** (see Subsection 2.1.2). In many places, the geostrophic current is a good approximation to the time-average ocean velocity (see Subsection 2.2.1). Over the globe, sea surface height varies over 3–4 m due to ocean currents.

Sea surface height is defined with respect to the **geoid**, which is the surface of equal gravitational potential, meaning that Earth's gravitational force is perpendicular to the geoid. Sea level would assume the shape of the geoid in the absence of ocean currents and other forces, such as those that generate the tide. The geoid itself is a complex surface that has peaks and troughs of O(10–100) m because Earth's gravity field varies from place to place. SSH variability is relatively easy to measure to an accuracy of a few cm, because the geoid is generally constant on the decadal timescales of satellite missions (though tides must be eliminated, among other corrections). Uncertainties in the exact shape of the geoid have limited the ability of altimetry to estimate long-term averages of SSH which are most relevant to this book, however. Uncertainties are particularly large in determining geoid features on a scale of 100 km or less. Only in the past decade have such uncertainties also been reduced to a few cm.

1.1.3 Platforms to Observe the Ocean

Understanding how oceanographers glean information about the general circulation does not rely simply on understanding oceanographic sensors. We must also appreciate the different ways that sensors are deployed in practice, and there is a tremendous diversity of ocean observing *platforms* on which to deploy sensors. The most important platforms are as follows.

The **ship** is the traditional platform for oceanographic measurements. The ship lowers instruments on a cable in order to measure how physical variables depend on depth at a given latitude and longitude. A **conductivity–temperature–depth (CTD)** instrument is commonly used to measure T, S, and p. CTD instruments are usually mounted on a "rosette" of water samplers arranged in a circle on a frame. The water sampler is called a "Niskin bottle" after its inventor, Shale Niskin, and is an open plastic tube that floods with water as the rosette sinks. A mechanism closes the tubes in order to capture water samples at separate depths. Chemical concentrations, including salinity (to calibrate the CTD conductivity measurements), and trace chemicals, such as oxygen and CFCs, can then be measured from the water sample in a laboratory, often onboard. Other instruments such as ADCPs can also be mounted on the rosette frame or attached to the ship's hull.

Ships can also take measurements while underway. Expendable instruments, such as an **expendable bathythermograph**, and more recently an expendable CTD, can be dropped off a ship and measure a vertical profile of temperature and salinity. The data are transmitted up a thin wire, typically several hundred meters long. The wire eventually breaks and the instrument is abandoned to fall to the seafloor. Many more T observations are taken with expendable bathythermographs than with CTDs, though the expendable bathythermograph has an inferior accuracy of only about 0.1°C and correcting for temperature biases can be challenging. Ships also sample meteorological variables such as sea-level pressure and wind speed, SST (often estimated from the ship's intake of engine-cooling water), and, less frequently, surface salinity. Until the mid-twentieth century, surface temperature measurements were made with a bucket lowered over the side, which is less accurate and has biases compared to more modern techniques.

Autonomous vehicles are devices that are released into the sea. They carry in-situ instruments to measure the properties of the water and deliver their observations to satellites from the sea surface, or acoustically to a data logger while submerged. Autonomous vehicles also provide important information about currents via successive fixes on their positions to yield their displacements over a known period of time.

Three important types of autonomous vehicle are drifters, floats, and gliders. There exist many other types of robotic underwater and surface vehicles with interesting characteristics, but they have been (so far) less relevant for studying the general circulation.

Drifters float at the sea surface and drift with the currents. They measure surface currents and SST, and sometimes salinity, atmospheric pressure, and wind. The Global Drifter Program coordinates drifter deployment and assembles the data. It maintains a fleet of 1250 drifters distributed throughout the global ocean.