Chapter 6

Temperature, Salinity, and Density

Insolation, evaporation, and rain influence the distribution of temperature and salinity at the ocean's surface. Changes in temperature and salinity lead to changes in the density of water at the surface, which can lead to convection and changes in the deeper circulation of the ocean. Once surface waters sink into the deeper ocean, they retain a distinctive relationship between temperature and salinity which helps oceanographers determine the source regions for subsurface water. In addition, changes in density lead to pressure changes inside the ocean and to changes in currents, which are driven by pressure gradients. For all these resons, we need to know the distribution of temperature, salinity, and density in the ocean.

Before discussing the distribution of temperature and salinity, let's first define what we mean by the terms, especially salinity.

6.1 Definition of Salinity

At the simplest level, salinity is the total amount of dissolved material in grams in one kilogram of sea water. Thus salinity is a dimensionless quantity. It has no units. The variability of dissolved salt is very small, and we must be very careful to define salinity in ways that are accurate and practical. To better understand the need for accuracy, look at figure 6.1. Notice that the range of salinity for most of the ocean's water is from 34.60 to 34.80 parts per thousand ($^{\circ}/_{\circ\circ}$), which is 200 parts per million. The variability in the deep North Pacific is even smaller, about 20 parts per million. If we want to classify water with different salinities, we need definitions and instruments accurate to about one part per million. Notice that the range of temperature is much larger, about 1°C, and temperature is easier to measure.

Writing a practical definition with useful accuracy is difficult (see Lewis, 1980, for the details), and various definitions have been used.





Figure 6.1 Histogram of temperature and salinity of ocean water colder than 4°C. Height is proportional to volume. Height of highest peak corresponds to a volume of 26 million cubic kilometers per bivariate class of 0.1° C and 0.01 (°/₀₀). From Worthington (1981).

A Simple Definition Originally salinity was defined to be the "Total amount of dissolved material in grams in one kilogram of sea water." This is not useful because the dissolved material is almost impossible to measure in practice. For example, how do we measure volatile material like gasses? Nor can we evaporate sea-water to dryness because chlorides are lost in the last stages of drying (Sverdrup, Johnson, and Fleming, 1942: 50).

A More Complete Definition To avoid these difficulties, the International Council for the Exploration of the Sea set up a commission in 1889 which recommended that salinity be defined as the "Total amount of solid materials in grams dissolved in one kilogram of sea water when all the carbonate has been converted to oxide, the bromine and iodine replaced by chlorine and all organic matter completely oxidized." The definition was published in 1902. This is useful but difficult to use routinely.

Salinity Based on Chlorinity Because the above definition was difficult to implement in practice, because salinity is directly proportional to the amount of chlorine in sea water, and because chlorine can be measured accurately by a simple chemical analysis, salinity S was redefined using chlorinity:

$$S(^{\circ}/_{\circ\circ}) = 0.03 + 1.805 Cl(^{\circ}/_{\circ\circ}) \tag{6.1}$$

where *chlorinity* Cl is defined as "the mass of silver required to precipitate completely the halogens in 0.328 523 4 kg of the sea-water sample."

As more and more accurate measurements were made, (6.1) turned out to be too inaccurate. In 1964 UNESCO and other international organizations appointed a Joint Panel on Oceanographic Tables and Standards to produce a more accurate definition. The Joint Panel recommended in 1966 (Wooster, Lee, and Dietrich, 1969) that salinity and clorinity be related using:

$$S(^{\circ}/_{\circ\circ}) = 1.80655Cl(^{\circ}/_{\circ\circ}) \tag{6.2}$$

This is the same as (6.1) for S = 35.

Salinity Based on Conductivity At the same time (6.2) was adopted, oceanographers had began using conductivity meters to measure salinity. The meters were very precise and relatively easy to use compared with the chemical techniques used to measure chlorinity. As a result, the Joint Panel also recommended that salinity be related to conductivity of sea water using:

$$S = -0.08996 + 28.29729 R_{15} + 12.80832 R_{15}^2 -10.67869 R_{15}^3 + 5.98624 R_{15}^4 - 1.32311 R_{15}^5$$
(6.3a)

$$R_{15} = C(S, 15, 0) / C(35, 15, 0)$$
(6.3b)

where C(S, 15, 0) is the conductivity of the sea-water sample at 15° C and atmospheric pressure, having a salinity S derived from (6.4), and C(35, 15, 0) is the conductivity of standard "Copenhagen" sea water. Millero (1996) points out that (6.3) is not a new definition of salinity, it merely gives chlorinity as a function of conductivity of seawater relative to standard seawater.

Practical Salinity Scale of 1978 By the early 1970s, accurate conductivity meters could be deployed from ships to measure conductivity at depth. The need to re-evaluate the salinity scale led the Joint Panel to recommend in 1978 that salinity be defined using only conductivity, breaking the link with chlorinity. All water samples with the same conductivity ratio have the same salinity.

The Practical Salinity Scale of 1978 is now the official definition:

$$S = 0.0080 - 0.1692 R_T^{1/2} + 25.3851 R_T + 14.0941 R_T^{3/2} - 7.0261 R_T^2 + 2.7081 R_T^{5/2} + \Delta S$$
(6.4a)
$$R_T = C(S, T, 0) / C(KCl, T, 0)$$
(6.4b)

$$\Delta S = \left[\frac{(T-15)}{1+0.0162(T-15)}\right] + 0.0005 - 0.0056 R_T^{1/2} - 0.0066 R_T - 0.0375 R_T^{3/2} + 0.636 R_T^2 - 0.0144 R_T^{5/2} 2 \le S \le 42$$
(6.4c)

where C(S, T, 0) is the conductivity of the sea-water sample at temperature T and standard atmospheric pressure, and C(KCl, T, 0) is the conductivity of the standard potassium chloride (KCl) solution at temperature T and standard atmospheric pressure. The standard KCl solution contains a mass of 32.4356 grams of KCl in a mass of 1.000 000 kg of solution. An extension of (6.4) gives salinity at any pressure (see Millero 1996: 72).

	Ion		Atoms
55.3%	Chlorine	55.3%	Chlorine
30.8%	Sodium	30.8%	Sodium
7.7%	Sulfate	3.7%	Magnesium
3.7%	Magnesium	2.6%	Sulfur
1.2%	Calcium	1.2%	Calcium
1.1%	Potassium	1.1%	Potassium

Table 6.1 Major Constituents of Sea Water

Comments The various definitions of salinity work well because the ratios of the various ions in sea water are nearly independent of salinity and location in the ocean (Table 6.1). Only very fresh waters, such as are found in estuaries, have significantly different ratios. The result is based on Dittmar's (1884) chemical analysis of 77 samples of sea water collected by the *Challenger* Expedition and further studies by Carritt and Carpenter (1958).

The importance of this result cannot be over emphasized, as upon it depends the validity of the chlorinity: salinity: density relationships and, hence, the accuracy of all conclusions based on the distribution of density where the latter is determined by chemical or indirect physical methods such as electrical conductivity...—Sverdrup, Johnson, Fleming (1942).

The relationship between conductivity and salinity has an accuracy of around ± 0.003 in salinity. The very small error is caused by variations in constituents such as SiO₂ which cause small changes in density but no change in conductivity.

Instruments for measuring salinity are calibrated using Normal Standard Seawater (P-series). The standard water is made from large samples of water from the north Atlantic carefully diluted to S = 35 which is distributed in 275ml sealed glass ampoules. Each is labelled for its conductivity ratio and salinity according to the Practical Salinity Scale 1978 and distributed worldwide by Ocean Scientific International in England since 1989. Each sample is carefully calibrated using the standard KCl solution.

6.2 Definition of Temperature

Many physical processes depend on temperature; and a few can be used to define absolute temperature T. The unit of T is the kelvin, which has the symbol K. The fundamental processes used for defining an absolute temperature scale over the range of temperatures found in the ocean include (Soulen and Fogle, 1997): 1) the gas laws relating pressure to temperature of an ideal gas with corrections for the density of the gas; and 2) the voltage noise of a resistance R.

The measurement of temperature using an absolute scale is difficult and the measurement is usually made by national standards laboratories. The absolute measurements are used to define a practical temperature scale based on the temperature of a few fixed points and interpolating devices which are calibrated at the fixed points.

For temperatures commonly found in the ocean, the interpolating device is a platinum-resistance thermometer. It consists of a loosly wound, strain-free, pure



Average Sea-Surface Temperature for July

Figure 6.3 Mean sea-surface temperature calculated from the optimal interpolation technique (Reynolds and Smith, 1995) using ship reports and AVHRR measurements of temperature. Contour interval is 1°C with heavy contours every 5°C. Shaded areas exceed 29°C.

platinum wire whose resistance is a function of temperature. It is calibrated at fixed points between the triple point of equilibrium hydrogen at 13.8033 K and the freezing point of silver at 961.78 K, including the triple point of water at 0.060°C, the melting point of Gallium at 29.7646°C, and the freezing point of Indium at 156.5985°C (Preston-Thomas, 1990). The triple point of water is the temperature at which ice, water, and water vapor are in equilibrium. The temperature scale in kelvin T is related to the temperature scale in degrees Celsius $t/^{\circ}$ C by:

$$t [^{\circ}C] = T [K] - 273.15$$
 (6.5)

The practical temperature scale was revised in 1887, 1927, 1948, 1968, and 1990 as more accurate determinations of absolute temperature become accepted. The most recent scale is the International Temperature Scale of 1990 (ITS-90). It differs slightly from the International Practical Temperature Scale of 1968 IPTS-68. At 0°C they are the same, and above 0°C ITS-90 is slightly cooler. $t_{90} - t_{68} = -0.002$ at 10°C, -0.005 at 20°C, -0.007 at 30°C and -0.010 at 40°C.

Notice that while oceanographers use thermometers calibrated with an accuracy of a millidegree, which is 0.001°C, the temperature scale itself has uncertainties of a few millidegrees.

6.3 Geographical Distribution of Surface Temperature and Salinity The distribution of temperature at the sea surface tends to be *zonal*, that is, it is independent of longitude (Figure 6.3). Warmest water is near the equator, coldest water is near the poles. The deviations from zonal are small. Equatorward of 40° , cooler waters tend to be on the eastern side of the basin. North of this latitude, cooler waters tend to be on the western side.

The *anomalies* of sea-surface temperature, the deviation from a long term average, are small, less than 1.5° C except in the equatorial Pacific where the deviations can be 3° C (Figure 6.4).

The annual range of sea-surface temperature is highest at mid-latitudes, especially on the western side of the ocean (Figure 6.5). In the west, cold air blows off the continents in winter and cools the ocean. The cooling dominates the heat budget. In the tropics the temperature range is mostly less than 2° C.

The distribution of sea-surface salinity also tends to be zonal. The saltiest waters are at mid-latitudes where evaporation is high. Less salty waters are near



Sea-Surface Temperature Anomalies for January 1996

Figure 6.4 Sea-surface temperature anomaly for January 1995 relative to mean temperature shown in figure 6.3 using data published by Reynolds and Smith (1995) in the *Climate Diagnostics Bulletin* for February 1995. Contour interval is 1°C; shaded areas are positive.



Annual Range of Sea-Surface Temperature

Figure 6.5 Annual range of sea-surface temperature in $^{\circ}C$ calculated from the Reynolds and Smith (1995) mean sea-surface temperature data set. Contour interval is $1^{\circ}C$ with heavy contours at $4^{\circ}C$ and $8^{\circ}C$. Shaded areas exceed $8^{\circ}C$.



Figure 6.6 Mean sea-surface salinity. Contour interval is 0.25 psu. Shaded areas exceed 36 psu. (From Levitus, 1982).

the equator where rain freshens the surface water, and at high latitudes where melted sea ice freshens the surface waters (Figure 6.6). The zonal (east-west) average of salinity shows a close correlation between salinity and evaporation minus precipitation plus river input (Figure 6.7).

Because many large rivers drain into the Atlantic and the Arctic Sea, why is the Atlantic saltier than the Pacific? Broecker (1997) showed that 0.32 Sv of the water evaporated from the Atlantic does not fall as rain on land. Instead, it is carried by winds into the Pacific (Figure 6.8). Broecker points out that the quantity is small, equivalent to a little more than the flow in the Amazon River,



Figure 6.7 Zonal average of sea-surface salinity calculated for all oceans from Levitus (1982) and the difference between evaporation and precipitation (E - P) calculated from data shown in Figure 5.14.

but "were this flux not compensated by an exchange of more salty Atlantic waters for less salty Pacific waters, the salinity of the entire Atlantic would rise about 1 gram per liter per millennium."

Mean Temperature and Salinity of the Ocean The mean temperature of the ocean's waters is: $T = 3.5^{\circ}C$; and the mean salinity is S = 34.7. The distribution about the mean is small: 50% of the water is in the range:



Figure 6.8 Water transported by the atmosphere into and out of the Atlantic. Basins draining into the Atlantic are black, deserts are white, and other drainage basins are shaded. Arrows give direction of water transport by the atmosphere, and values are in Sverdrups. Bold numbers give the net transport for the Atlantic. Averall, the Atlantic loses 0.32 Sv, an amount approximately equal to the flow in the Amazon River. From Broecker (1997).

$$1.3^{\circ}C < T < 3.8^{\circ}C$$

 $34.6 < S < 34.8$

6.4 The Oceanic Mixed Layer

Wind blowing on the ocean stirs the upper layers leading to a thin *mixed layer* at the sea surface having constant temperature and salinity from the surface down to a depth where the values differ from those at the surface. The magnitude of the difference is arbitrary, but typically the temperature at the bottom of the layer must be no more than $0.02-0.1^{\circ}$ colder than at the surface. Note that the both temperature and salinity must be constant in the mixed layer. We will see later that mean velocity can vary with depth in the mixed layer.

The mixed layer is roughly 10–200 m thick over most of the tropical and midlatitude belts (Figure 6.9 upper). The mixed layer also tends to be saltier than the deeper layers except at high latitudes (Figure 6.9 lower). Below the mixed layer, water temperature rapidly decreases with depth. The range of depths where the rate of change, the gradient of temperature, is large is called the *thermocline*. Because density is closely related to temperature, the thermocline also tends to be the layer where density gradient is greatest, the *pyncocline*.

The depth and temperature of the mixed layer varies from day to day and from season to season in response to two processes:

- 1. Heat fluxes through the surface heat and cool the surface waters. Changes in temperature change the density contrast between the mixed layer and deeper waters. The greater the contrast, the more work is needed to mix the layer downward and visa versa.
- 2. Turbulence in the mixed layer provides the mechanical work necessary to mix heat downward. The turbulence depends on the wind speed and on the intensity of breaking waves. Turbulence mixes water in the layer, and it mixes the water in the layer with water in the thermocline.

The mid-latitude mixed layer is thinnest in late summer when winds are weak, and sunlight warms the surface layer. At times, the heating is so strong, and the winds so weak, that the layer is only a few meters thick. In Fall, early storms mix the heat down into the ocean thickening the mixed layer, but little heat is lost. In Winter, heat is lost, and the mixed layer continues to thicken, becoming thickest in late winter. In Spring, winds weaken, sunlight increases, and a new mixed layer forms (Figure 6.10).

The mixed layer rarely extends below two hundred meters. Below the upper two hundred meters is a permanent thermocline that merges with the cold, deep waters of the ocean's interior.

6.5 Potential Temperature, Potential Density, and Sigma

As water sinks and flows into the deep ocean, it can move far from its original source at the surface. To trace the movement of water in the deep ocean, we must compare temperature at one depth with temperature at another. This



Figure 6.9 **Upper**: Typical mean temperature profiles in the open ocean. **Lower**: Typical mean salinity profiles in the open ocean. (From Pickard and Emery, 1990).

is possible, but difficult. As pressure increases, water is compressed, and the compression does work on the water. This causes the water to warm. To understand the warming, consider a cube containing a fixed mass of water. As the cube sinks, its sides move inward as the cube is compressed. Recalling that work is force times distance, we note that work is the distance the side moves times the force exerted on the side by pressure. The heating is small but noticeable compared with the small changes of the temperature of the adjacent water (Figure 6.11).

Potential Temperature To avoid calculating temperature changes due to compressibility of water, oceanographers (and meteorologists who have the same problem in the atmosphere) use the concept of potential temperature. *Potential temperature* is defined as the temperature of a parcel of water at the sea surface after it has been raised adiabically from some depth in the ocean. Raising the parcel *adiabically* means that it is raised in an insulated container so it does not exchange heat with its surroundings. Of course, the parcel is not actually brought to the surface. Potential temperature is calculated from the



Figure 6.10 Growth and decay of the seasonal thermocline at Ocean Station "Pappa" at 50° N, 145°W in the North Pacific. (From Pickard and Emery, 1990).

temperature in the water at depth, the *in situ* temperature.

Density and sigma-t Density is another important property of sea water. Less dense water floats on more dense water, and if we wish to determine how water can move within the ocean, we need to be able to calculate the density of water with an accuracy of a few parts per million.

Absolute Density of water can only be measured in spacial laboratories, and only with difficulty. The best accuracy is $1 : 2.5 \times 10^5 = 4$ parts per million.

To avoid the difficulty of working with absolute density, oceanographers use density relative to density of pure water. Density $\rho(s,t,p)$ is now defined using Standard Mean Ocean Water of known isotopic composition, assuming saturation of dissolved atmospheric gasses. Here s, t, p refers to salinity, temperature, and pressure.

In practice, density is not measured, it is calculated from *in situ* measurements of pressure, temperature, and conductivity using the equation of state



Figure 6.11 Profiles of *in situ* and potential temperature and density in the Mindinao Trench in the Pacific: (a,b) vertical profiles, (c,d) vertical sections. (From Pickard and Emery, 1990).

for sea water. This can be done with an accuracy of two parts per million.

Density of water at the sea surface is typically 1027 kg/m³. For simplification, physical oceanographers often quote only the last 2 digits of the density, a quantity called the *density anomaly* or Sigma (s,t,p):

$$\sigma(s, t, p) = \rho(s, t, p) - 1000 \text{ kg/m}^3$$
(6.6)

 $\sigma(s, t, p)$ is typically 27.00 kg/m³. The Working Group on Symbols in Oceanography recommends that σ be replaced by γ because σ was originally defined relative to pure water and it was dimensionless. Here, however, we will follow common practice and use σ .

For studies of the surface layers of the ocean, compressibility can be ignored, and a new quantity sigma-t (written σ_t) is used:

$$\sigma_t = \sigma(s, t, 0) \tag{6.7}$$

This is the density anomaly of a water sample when the total pressure on it has been reduced to atmospheric pressure (*i.e.* zero water pressure), but the temperature and salinity are *in situ* values.

Potential Density For studies of processes deeper within the ocean, compressibility cannot be ignored. Because changes in pressure primarily influence the temperature of the water, the influence of pressure can be removed, to a first approximation, by using the *potential density*.

Potential density is the density a parcel of water would have if it were raised adiabatically to the surface. The potential density anomaly of such a sample is σ_{θ} .

$$\sigma_{\theta} = \sigma(s, \theta, 0) \tag{6.8}$$

Potential density is useful because it removes the primary influence of pressure on density. This allows us to compare density of water samples from different depths. It is also useful because water tends to flow along surfaces of constant potential density.

 $\sigma(\theta)$ is not useful for comparing density of water at great depths because the relation between density and temperature and salinity is non-linear. For example, two water samples having the same density but different temperature and salinity at a depth of four kilometers can have noticeably different density when moved adiabatically to the sea surface. Thus the use of $\sigma(\theta)$ can lead to an apparent decrease of density with depth (Figure 6.12) although we know that this is not possible because such a column of water would be unstable.

If samples from great depths are compared, it is better to use sigma values for a depth of 4 km:

$$\sigma_4 = \sigma(s, \theta, 4000) \tag{6.9}$$

where σ_4 is the density of a parcel of water lowered adiabically to a depth of 4 km. In general, oceanographers use σ_n , where n is pressure in decibars divided by 1000.

Equation of state of sea water Density of sea water is rarely measured. Density is calculated from measurements of temperature, conductivity, or salinity, and pressure using the equation of state of sea water. The equation of state is an equation relating density to temperature, salinity, and pressure.

The equation is derived by fitting curves through laboratory measurements of density as a function of temperature, pressure, and salinity, chlorinity, or conductivity. The International Equation of State (1980) published by the Joint Panel on Oceanographic Tables and Standards (1981) is now used. See also Millero and Poisson (1981) and Millero et al (1980). The equation has an accuracy of 10 parts per million, which is 0.01 units of $\sigma(\theta)$.



Figure 6.12 Vertical sections of density in the western Atlantic. Depth scale changes at 1000 m depth. Hydrographic stations are noted by dots. **Upper**: Sigma- θ , showing an apparent density inversion below 3,000 m. **Lower**: Sigma(4) showing continuous increase in density with depth. From Lynn and Reid, 1968.

I have not actually written out the equation of state because it consists of three polynomials with 41 constants (JPOTS, 1991).

Accuracy of Temperature, Salinity, and Density If we are to distinguish between different water masses in the ocean, and if the total range of temperature and salinity is as small as the range in Figure 6.1, then we must determine temperature, salinity, and pressure very carefully. We will need an accuracy of a few parts per million.

Such accuracy can be achieved only if all quantities are carefully defined, if all measurements are made with great care, and if all instruments are carefully calibrated. All who use hydrographic data pay careful attention to the directions in the *Processing of Oceanographic Station Data* (JPOTS, 1991) published by UNESCO. The book contains internationally accepted definitions of primary variables such as temperature and salinity and methods for the measuring the primary variables. It also describes accepted methods for calculating quantities derived from primary variables, such as potential temperature, density, and stability. Part of the material in this chapter are based largely on information contained in *Processing of Oceanographic Station Data*.

6.6 Measurement of Temperature

Temperature in the ocean has been measured many ways. Thermistors and mercury thermometers are commonly used on ships and buoys. These are calibrated in the laboratory before being used, and after use if possible, using mercury or platinum thermometers with accuracy traceable to national standards laboratories. Infrared radiometers are used in space to observe the surface temperature of the ocean.

Mercury Thermometer This is the most widely used, non-electronic thermometer. It is used in buckets dropped over the side of a ship to measure the temperature of surface waters, on Nansen bottles to measure subsea temperatures, and in the laboratory to calibrate other thermometers. Accuracy is about $\pm 0.001^{\circ}$ C with careful calibration.

One very important mercury thermometer is the reversing thermometer (Figure 6.13) carried on Nansen bottles, which are described in the next section. It is a thermometer that has a constriction in the mercury capillary that causes the thread of mercury to break at a precisely determined point when the thermometer is turned upside down. The thermometer is lowered deep into the ocean in the normal position; and it is allowed to come to equilibrium with the water. Mercury expands into the capillary, and the amount of mercury in the capillary is proportional to temperature. The thermometer is then flipped upside down, the thread of mercury breaks trapping the mercury in the capilalry, and the thermometer is brought back. The mercury in the capillary of the reversed thermometer is read on deck along with the temperature of a normal thermometer, which gives the temperature at which the reversed thermometer is read. The two readings give the temperature of the water at the depth where the thermometer was reversed.

The reversing thermometer is carried inside a glass tube which protects the thermometer from the ocean's pressure because high pressure can squeeze additional mercury into the capillary. If the thermometer is unprotected, the apparent temperature read on deck is proportional to temperature and pressure at the depth where the thermometer was flipped. A pair of protected and unprotected thermometers gives temperature and pressure of the water at the depth the thermometer was reversed.

Platinum Resistance Thermometer This is the standard for temperature. It is used by national standards laboratories to interpolate between defined points on the practical temperature scale. It is also used on instruments deployed from ships because it can be read electronically with great accuracy.



Figure 6.13 Left: Protected and unprotected reversing thermometers is set position, before reversal. **Right**: The constricted part of the capillary in set and reversed positions (From von Arx, 1962).

Thermistor A thermistor is a resistor made of semiconductors having resistance that varies rapidly and predictably with temperature. The themistor is widely used on moored instruments and on instruments deployed from ships. It has high resolution and an accuracy of about $\pm 0.001^{\circ}$ C when carefully calibrated.

Bucket temperatures The temperature of surface waters have been routinely measured at sea by putting a mercury thermometer into a bucket which is lowered into the water, letting it sit at a depth of about a meter for a few minutes until the thermometer comes to equilibrium, then bringing it aboard and reading the temperature before water in the bucket has time to change temperature. The accuracy is around 0.1°C.

Ship Injection Temperature The temperature of the water drawn into the ship to cool the engines has been recorded routinely for decades, and the observations have been collected in archives. These recorded values of temperature are called injection temperatures. Errors are due to ship's structure warming water before it is recorded. This happens when the temperature recorder is not placed close to the point on the hull where water is brought in. Accuracy is $0.5^{\circ}-1^{\circ}C$.

Advanced Very High Resolution Radiometer The most commonly used instrument to measure sea-surface temperature from space is the Advanced Very High Resolution Radiometer AVHRR. The instrument has been carried on all polar-orbiting meteorological satellites operated by NOAA since Tiros-N was launched in 1978.

The instrument was originally designed to measure cloud temperatures and hence cloud height. The instrument had, however, sufficient accuracy and precision that it was soon used to measure regional and global temperature patterns at the sea surface.

The instrument is a radiometer that converts observed radiation into electrical signals. It includes a mirror that scans from side to side across the subsatellite track and reflects radiance from the ground into a telescope, a telescope that focuses the radiance on detectors, detectors sensitive to different wavelengths that convert the radiance at those wavelengths into electrical signals, and electronic circuitry to digitize and store the radiance values. The instruments observes a 2700-km wide swath centered on the subsatellite track. Each observation along the scan is from a pixel that is roughly one kilometer in diameter near the center of the scan and that increases in size with distance from the subsatellite track.

The radiometers measures infrared radiation emitted from the surface in five wavelength bands: three infrared bands: $3.55-3.99 \ \mu\text{m}$, $10.3-11.3 \ \mu\text{m}$, and $11.5-12.5 \ \mu\text{m}$; a near-infrared band at $0.725-1.10 \ \mu\text{m}$; and a visible-light band at $0.55-0.90 \ \mu\text{m}$. All infrared bands include radiation emitted from the sea and from water vapor in the air along the path from the satellite to the ground. The $3.7 \ \mu\text{m}$ band is least sensitive to water vapor and other errors, but it works only at night because sunlight has radiance in this band. The two longest wavelength bands at $10.8 \ \mu\text{m}$ and $12.0 \ \mu\text{m}$ are used to observe sea-surface temperature and water vapor along the path in daylight.

Data with 1-km resolution are transmitted directly to ground stations that view the satellite as it passes the station. This is the Local Area Coverage mode. Data are also averaged to produce observations from 4×4 km pixels. These data are stored on tape recorders and later transmitted to NOAA receiving stations. This is the Global Area Coverage mode.

The swath width is sufficiently wide that the satellite views the entire earth twice per day, at approximately 09:00 AM and 9:00 PM local time. Areas at high latitudes may be observed as often as eight or more times per day.

Errors in the measurement of sea-surface temperature are due to:

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Figure 6.14 The influence of clouds on infrared observations. Left: The standard deviation of the radiance from small, partly cloudy areas each containing 64 pixels. The feet of the arch-like distribution of points are the sea-surface and cloud-top temperatures. (After Coakley and Bretherton (1982). **Right:** The maximum difference between local values of $T_{11} - T_{3.7}$ and the local mean values of the same quantity. Values inside the dashed box indicate cloud-free pixels. T_{11} and $T_{3.7}$ are the apparent temperatures at 11.0 and 3.7 μ m (data from K. Kelly). From Stewart (1985).

- 1. Unresolved or undetected clouds: Large, thick clouds are obvious in the images of water temperature Thin clouds such as low stratus and high cirrus produce much small errors that are difficult or almost impossible to detect. Clouds smaller in diameter than 1 km, such as trade-wind cumuli, are also difficult to detect. Special techniques have been developed for detecting small clouds (Figure 6.14).
- 2. Water vapor, which absorbs part of the energy radiated from the sea surface: Water vapor reduces the apparent temperature of the sea surface. The influence is different in the 10.8 μ m and 12.0 μ m channels, allowing the difference in the two signals to be used to reduce the error.
- 3. Aerosols, which absorb infrared radiation. They radiate at temperatures found high in the atmosphere. Stratospheric aerosols generated by volcanic eruptions can lower the observed temperatures by up to a few degrees Celcius, and dust particles carried over the Atlantic from Saharan dust storms can also cause errors.
- 4. Instrument noise tends to be small, limiting the temperature resolution in local-area images.
- 5. Skin temperature errors. The infrared radiation seen by the instrument comes from a layer at the sea surface that is only a few micrometers thick. The temperature in this layer is not the same as temperature a meter below the sea surface. They can differ by several degrees when winds are light (Emery and Schussel, 1989).

Maps of temperature processed from Local Area Coverage of cloud-free regions show variations of temperature with a precision of 0.1° C. These maps are useful for observing local phenomena including patterns produced by local currents. Figure 10.17 shows such patterns off the California coast.

Global maps are much more difficult to produce. The U.S. Naval Oceanographic Office receives the global AVHRR data directly from NOAA's National Environmental Satellite, Data and Information Service in near-real time each day. The data are carefully processed to remove the influence of clouds, water vapor, aerosols, and other sources of error. Data are then used to produce global maps between $\pm 70^{\circ}$ with an accuracy of $\pm 0.6^{\circ}$ C (May et al 1998). The maps of sea-surface temperature are sent to the U.S. Navy and to NOAA's National Centers for Environmental Prediction. In addition, the office produces daily 100-km global and 14-km regional maps of temperature.

Global Maps of Sea-Surface Temperature Global, monthly maps of surface temperature are produced by the National Centers for Environmental Prediction using Reynolds' (1988, 1993, 1994) optimal-interpolation method. The technique blends ship and buoy measurements of sea-surface temperature with AVHRR data processed by the Naval Oceanographic Office in 1° areas for a month. Essentially, AVHRR data are interpolated between buoy and ship reports using previous information about the temperature field. Overall accuracy ranges from approximately $\pm 0.3^{\circ}$ C in the tropics to $\pm 0.5^{\circ}$ C near western boundary currents in the northern hemisphere where temperature gradients are large. Maps are available from November 1981. Figures 6.3–6.5 were made by NOAA using Reynolds' technique.

Maps of mean temperature have also been made from COADS data. The data are poorly distributed in time and space except for some areas of the northern hemisphere. In addition, Reynolds and Smith (1994) found that ship temperature data had errors twice as large as temperature errors in data from buoys and AVHRR. Thus, space data processed by Reynolds are more accurate, and better distributed than COADS.

Anomalies of sea-surface temperature are calculated using mean sea-surface temperature from the period 1950–1979 calculated from COADS supplemented with four years of satellite data 1982–1985 (Reynolds and Smith, 1995).

6.7 Measurement of Conductivity

Measurements of conductivity can be made using electrodes, but electrodes tend to drift due to electrochemical processes. Remember, two different metals dipped into a conducting solution make a battery.

Measurements are usually made using *induction*. The sea water forms one side of a transformer, and the current induced in the transformer coils depends on conductivity of sea water (Figure 6.15). The technique eliminates electrochemical drifts. The best measurements of salinity from conductivity give salinity with an accuracy of ± 0.005 psu.

Salinity is sometimes measured using chemical titration of the water sample with silver salts. The best measurements of salinity from titration give salinity with an accuracy of ± 0.02 psu.

Instruments for measuring salinity are calibrated using standard seawater.



Figure 6.15 The induction method measurement of electrical conductivity uses a sea-water loop which couples two transformer loops T_1 and T_2 . The induced voltage in the second loop is proportional to conductivity R_w of the water. From Dietrich et al. 1980.

Long-term studies of accuracy sometimes use data from measurements of deep water masses of known, stable, salinity. For example, Saunders (1986) noted that temperature is very accurately related to salinity for a large volume of water contained in the deep basin of the northwest Atlantic under the Mediterranean outflow. He used the consistency of measurements of temperature and salinity made at many hydrographic stations in the area to estimate the accuracy of temperature, salinity and oxygen measurements. He concluded that the most careful measurements made since 1970 have an accuracy of 0.005 psu for salinity and 0.005°C for temperature. The largest source of salinity error was the error in determination of the standard water used for calibrating the salinity measurements.

Gouretski and Jancke (1995) estimated accuracy of salinity measurements as a function of time. They used high quality measurements from 16,000 hydrographic stations in the south Atlantic from 1912 to 1991 to produce a consistent



Figure 6.16. Standard deviation of salinity measurements at depths below 1500 m in the South Atlantic from 1920 to 1993. Each point is the average of data collected for the decade centered on the point. The value for 1995 is an estimate of the accuracy of recent measurements. From Table 1 of Gouretski and Jancke (1995).



Figure 6.17. Results from a test of a Sea-Bird Electronics 911 Plus CTD made in North Atlantic Deep Water in 1992 (From Sea-Bird Electronics, 1992).

data set. They estimated accuracy by plotting salinity as a function of temperature using all data collected below 1500 m in twelve regions for each decade from 1920 to 1990. For small ranges of temperature, salinity was a linear function of temperature. The accuracy of the salinity measurement was calculated from the standard deviation of the salinity values relative to the best-fitting straight lines through the data. A plot of accuracy as a function of time since 1920 shows consistent improvement in accuracy since 1950 (Figure 6.16). Recent measurements of salinity are the most accurate. The standard deviation of modern salinity data collected from all areas in the South Atlantic from 1970 to 1993 adjusted as described by Gouretski and Jancke (1995) was 0.0033 psu. More recent instruments such as the Sea-Bird Electronics Model 911 Plus have an accuracy of better than 0.005 psu without adjustments. A careful comparison of salinity measured at $43^{\circ}10$ 'N $14^{\circ}4.5$ 'W by the 911 Plus with historic data collected by Saunders (1986) gives an accuracy of 0.002 psu (Figure 6.17).

6.8 Measurement of Pressure

Pressure is routinely measured by many different types of instruments. The SI unit of pressure is the pascal (Pa), but oceanographers normally report pressure in decibars (dbar), where:

$$1 \text{ dbar} = 10^4 \text{ Pa}$$
 (6.10)

because the pressure in decibars is almost exactly equal to the depth in meters. Thus 1000 dbar is the pressure at a depth of about 1000 m.

Strain Gage This is the simplest and cheapest instrument, and it is widely used. Accuracy is about $\pm 1\%$.

Vibratron Much more accurate measurements of pressure can be made by measuring the natural frequency of a vibrating tungsten wire stretched in a magnetic field between diaphrams closing the ends of a cylinder. Presure distorts the diaphram, which changes the tension on the wire and its frequency. The frequency can be measured from the changing voltage induced as the wire vibrates in the magnetic field. Accuracy is about $\pm 0.1\%$, or better when temperature controlled. Precision is 100–1000 times better than accuracy. The instrument is useful for detecting small changes in pressure at great depths. Snodgrass (1968) obtained ± 0.8 mm precision in 3 km depth.

Quartz crystal Very accurate measurements of pressure can also be made by measuring the natural frequancy of a quartz crystal cut for minimum temperature dependence. The best accuracy is obtained when the temperature of the crystal is held constant. The accuracy is $\pm 0.015\%$; and precision is $\pm 0.001\%$ resolution (of full-scale values).

Quartz Bourdon Gage has accuracy and stability comparable to quartz crystals. It too is used for long-term measurements of pressure in the deep sea.

6.9 Measurement of Temperature and Salinity with Depth

Temperature, salinity, and pressure are measured as a function of depth using various instruments or techniques, and density is calculated from the measurements.

Bathythermograph (BT) is a mechanical device that measures temperature vs depth on a smoked glass slide. The device was widely used to map the thermal structure of the upper ocean, including the depth of the mixed layer before being replaced by the expendable bethythermograph in the 1970s.

Expendable Bathythermograph (XBT) is an electronic device that measures temperature vs depth using a thermistor on a free-falling streamlined weight. The thermistor is connected to an ohm-meter on the ship by a thin copper wire that is spooled out from the sinking weight and from the moving ship. The XBT is now the most widely used instrument for measuring the thermal structure of the upper ocean. Approximately 65,000 are used each year.

- 1. Velocity of fall is constant, giving depth accuracy of $\pm 2\%$.
- 2. Temperature accuracy is $\pm 0.1^{\circ}$ C.
- 3. Vertical resolution is typically 65 cm.
- 4. Probes reach to depths of 200 m to 1830 m depending on model.

Nansen Bottles (Figure 6.18) were deployed from ships stopped at hydrographic stations. *Hydrographic stations* are places where oceanographers measure water properties from the surface to some depth, or to the bottom, using instruments lowered from a ship. Usually 20 bottles were attached at intervals of a few tens to hundreds of meters to a wire lowered over the side of the ship. The distribution with depth was selected so that most bottles are in the upper layers of the water column where the rate of change of temperature in the vertical is greatest. A protected reversing thermometer for measuring temperature



Figure 6.18 Left A CTD ready to be lowered over the side of a ship (From Davis, 1987). Right Nansen water bottles before (I), during (II), and after (III) reversing. Both instruments are shown at close to the same scale (From Dietrich et al. 1980).

was attached to each bottle along with an unprotected reversing thermometer for measuring depth. The bottle contains a tube with valves on each end to collect sea water at depth. Salinity was determined by laboratory analysis of water sample collected at depth.

After bottles had been attached to the wire and all had been lowered to their selected depths, a lead weight was sent down the wire. The weight tripped a mechanism on each bottle, and the bottle flipped over, reversing the thermometers, shutting the valves and trapping water in the tube, and releasing another weight. When all bottles had been tripped, the string of bottles was recovered. The deployment and retrieval typically took several hours.

CTD Mechanical instruments on Nansen bottles were replaced beginning in the 1960s by an electronic instrument, called a CTD, that measured conductivity, temperature, and depth (Figure 6.18). The measurements are recorded in digital form either within the instrument as it is lowered from a ship or on the ship. Temperature is usually measured by a thermistor; conductivity is measured by induction; pressure is measured by a quartz crystal. Modern instruments have accuracy summarized in Table 6.2.

Table 0.2 Summary of Measurement Recuracy				
Variable	Range	Best Accuracy		
Temperature	$42 \ ^{\circ}\mathrm{C}$	\pm 0.001 °C		
Salinity	1 psu	\pm 0.02 psu by titration		
		\pm 0.005 psu by conductivity		
Pressure	10,00 dbar	± 0.65 dbar		
Density	2 kg/m^3	$\pm 0.005 \text{ kg/m}^3$		
Equation of State		$\pm 0.005 \text{ kg/m}^3$		

Table 6.2 Summary of Measurement Accuracy

6.10 Measurements of Mixed-Layer Depth

The depth of the mixed layer is usually calculated from bathythermograph data. But, be careful using only temperature observations because in some areas the constant-temperature layer may be thicker than the constant-salinity layer. It is better to calculate mixed-layer depth using both temperature and salinity measurements from a CTD.

The depth of the mixed layer can also be calculated from temperatures observed from space, although this is not widely done. This technique is useful for studies of areas seldom visited by ships. It makes use of the thermal inertia of the mixed layer. For example, suppose we calculated the heat flux into or out of an oceanic area, say 5 ° on a side, for a period of a month. The change in heat must produce a change in temperature of the water. The temperature changes is proportional to the volume of water in close contact with the surface. This volume is the area times the mixed-layer depth. Thus, changes in temperature resulting from known changes in heat flux gives the depth of the mixed layer (Yan et al. 1991, 1995). The technique works only to the extent that advection can be neglected. If typical currents are 20 cm/s, which is 20 km/day, we might expect that the technique can be applied only to areas whose horizontal extent exceeds the distance water flows in the time period used in the calculation of heat change, or say roughly 25 days for 5° squares. The technique also assumes that heat advection out of the bottom of the mixed layer can be neglected.

6.11 Light in the ocean and absorption of light

Sunlight in the ocean is important for many reasons: It heats sea water, warming the surface layers; it provides energy required by phytoplankton; it is used for navigation by animals near the surface; and reflected subsurface light is used for mapping chlorophyll concentration from space.

Light in the ocean travels at a velocity equal to the velocity of light in a vacuum divided by the index of refraction (n), which is typically n = 1.33. Hence the velocity in water is about 2.25×10^8 m/s. Because light travels slower in water than in air, some light is reflected at the sea surface. For light shining straight down on the sea, the reflectivity is $(n - 1)^2/(n + 1)^2$. For seawater, the reflectivity is 0.02 = 2%. Hence most sunlight reaching the sea surface is transmitted into the sea, little is reflected. This means that sunlight incident on the ocean in the tropics is mostly absorbed below the sea surface, little sunlight is reflected back to the atmosphere.



Figure 6.19 Attenuation coefficient c and scattering coefficient b for pure water as a function of wavelength λ of the radiation (From Dietrich, et al. 1980).

The rate at which sunlight is attenuated determines the depth which is lighted and heated by the sun. Attenuation is due to absorption by pigments and scattering by molecules and particles. Attenuation depends on wavelength. Blue light is absorbed least, red light is absorbed most strongly. Attenuation per unit distance is proportional to the radiance or the irradiance of light:

$$\frac{dI}{dx} = -c I \tag{6.11}$$

where x is the distance along beam, c is an attenuation coefficient (Figure 6.19), and I is irradiance or irradiance.

Radiance is the power per unit area per solid angle. It is useful for describing the energy in a beam of light coming from a particular direction. Sometimes we want to know how much light reaches some depth in the ocean regardless of which direction it is going. In this case we use *irradiance*, which is the power per unit area of surface.

If the absorption coefficient is constant, the light intensity decreases exponentially with distance.

$$I_2 = I_1 \, \exp(-cx) \tag{6.12}$$

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Figure 6.20 Left: Attenuation of daylight in the ocean in % per meter as a function of wavelength. I: extremely pure ocean water; II: turbid tropical-subtropical water; III: mid-latitude water; 1-9: coastal waters of increasing turbidity. Incidence angle is 90° for the first three cases, 45° for the other cases. (From Jerlov, 1951). **Right:** Percentage of 465 nm light reaching indicated depths for the same types of water. (From Jerlov, 1968).

where I_1 is the original radiance or irradiance of light, and I_2 is the radiance or irradiance of light after absorption.

Clarity of Ocean Water Sea water in the middle of the ocean is very clear clearer than distilled water. These waters are a very deep, cobalt, blue—almost black. Thus the strong current which flows northward offshore of Japan carrying very clear water from the central Pacific into higher latitudes is known as the Black Current, or Kuroshio in Japanese. The clearest ocean water is called Type I waters by Jerlov (Figure 6.20). The water is so clear that 10% of the light transmitted below the sea surface reaches a depth of 90 m.

In the subtropics and mid-latitudes closer to the coast, sea water cantains more phytoplankton than the very clear central-ocean waters. Chlorophyll pigments in phytoplankton absorb light, and the plants themselves scatter light. Together, the processes change the color of the ocean as seen by observer looking downward into the sea. Very productive waters, those with high concentrations of phytoplankton, appear blue-green or green (Figure 6.21). On clear days the color can be observed from space. This allows ocean-color scanners, such as those on SeaWiFS, to map the distribution of phytoplankton over large areas.

As the concentration of phytoplankton increases, the depth where sunlight is absorbed in the ocean decreases. The more turbid tropical and mid-latitude waters are classified as type II and III waters by Jerlov (Figure 6.19). Thus the depth where sunlight warms the water depends on the productivity of the waters. This complicates the calculation of solar heating of the mixed layer.

Coastal waters are much less clear than waters offshore. These are the type 1–9 waters shown in figure 6.19. They contain pigments from land, sometimes called gelbstoffe, which just means yellow stuff, muddy water from rivers, and mud stirred up by waves in shallow water. Very little light penetrates more than a few meters into these waters.

Measurement of Chlorophyll from Space The color of the ocean, and hence the chlorophyll concentration in the upper layers of the ocean has been



Figure 6.21 Spectral reflectance of sea water observed from an aircraft flying at 305 m over waters of different colors in the Northwest Atlantic. The numerical values are the average chlorophyll concentration in the euphotic (sunlit) zone in units of mg/m^3 . The reflectance is for vertically polarized light observed at Brewster's angle of 53°. This angle minimizes reflected skylight and emphasizes the light from below the sea surface (From Clarke, Ewing, and Lorenzen, 1970).

measured by the Coastal Zone Color Scanner carried on the Nimbus-7 satellite launched in 1978 and by the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) carried on SeaStar, launched in 1997. The latter instrument measures upwelling radiance in eight wavelength bands between 412 nm and 856 nm.

Most of the upwelling radiance seen by the satellite comes from the atmosphere. Only about 10% comes from the sea surface. Both air molecules and aerosols scatter light; and very accurate techniques have been developed to remove the influence of the atmsophere.

The total radiance L_t received by an instrument in space is:

$$L_t(\lambda_i) = t(\lambda_i)L_W(\lambda_i) + L_r(\lambda_i) + L_a(\lambda_i)$$
(6.13)

where λ_i is the wavelength of the radiation in the band measured by the instrument, L_W is the radiance leaving the sea surface, L_r is radiance scattered by molecules, called the Rayleigh radiance, L_a is radiance scattered from aerosols, and t is the transmittance of the atmosphere. L_r can be calculated from theory; and L_a can be calculated from the amount of red light received at the instrument because very little red light is reflected from the water. Therefore L_W can be calculated from the radiance measured at the spacecraft.

Chlorophyll concentration in the water column is calculated from the ratio of L_W at two frequencies. Using data from the Coastal Zone Color Scanner, Gordon et al. (1983) proposed

$$C_{13} = 1.1298 \left[\frac{L_W(443)}{L_W(550)} \right]^{-1.71}$$
(6.14a)

$$C_{23} = 3.3266 \left[\frac{L_W(520)}{L_W(550)} \right]^{-2.40}$$
(6.14b)

where C is the chlorophyll concentration in the surface layers in mg pigment/m³, and $L_W(443), L_W(520), and L_W(550)$ are the radiances at wavelengths of 443, 520, and 550 nm. C_{13} is used when $C_{13} \leq 1.5 \text{ mg/m}^3$; otherwise C_{23} is used.

The technique is used to calculate chlorophyll concentration within a factor of 50% over a wide range of concentrations from 0.01 to 10 mg/m^3 .

6.12 Important Concepts

- 1. Density in the ocean is determined by temperature, salinity, and pressure.
- 2. Density changes in the ocean are very small, and studies of water masses and currents require density with an accuracy of 10 parts per million.
- 3. Density is not measured, it is calculated from measurements of temperature, salinity, and pressure using the equation of state of sea water.
- 4. Accurate calculations of density require accurate definitions of temperature and salinity and an accurate equation of state.
- 5. Salinity is difficult to define and to measure. To avoid the difficulty, oceanographers use conductivity instead of salinity. They measure conductivity and calculate density from temperature, conductivity, and pressure.
- 6. A mixed layer of constant temperature and salinity is usually found in the top 1–100 meters of the ocean. The depth is determined by wind speed and the flux of heat through the sea surface.
- 7. To compare temperature and density of water masses at different depths in the ocean, oceanographers use potential temperature and potential density which remove most of the influence of pressure on density.
- 8. Surface temperature of the ocean was usually measured at sea using bucket or injection temperatures. Global maps of temperature combine these observations with observations of infrared radiance from the sea surface measured by an AVHRR in space.
- 9. Temperature and conductivity are usually measured digitally as a function of pressure using a CTD. Before 1960–1970 the salinity and temperature were measured at a few depths using Nansen bottles lowered on a line from a ship. The bottles carried reversing thermometers which recorded temperature and depth and they returned a water sample from that depth which was used to determine salinity on board the ship.

- 10. Light is rapidly absorbed in the ocean. 95% of sunlight is absorbed in the upper 100 m of the clearest sea water. Sunlight rarely penetrates deeper than a few meters in turbid coastal waters.
- 11. Phytoplankton change the color of sea water, and the change in color can be observed from space. Water color is used to measure phytoplankton concentration from space.

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