Chapter 2

Temperature, salinity, density, and the oceanic pressure field

The ratios of the many components which make up the salt in the ocean are remarkably constant, and salinity, the total salt content of seawater, is a well-defined quantity. For a water sample of known temperature and pressure it can be determined by only one measurement, that of conductivity.

Today, the single most useful instrument for oceanographic measurements is the CTD, which stands for "Conductivity-Temperature-Depth". It is sometimes also known as the STD, which stands for "Salinity-Temperature-Depth"; but CTD is the more accurate description, because in both systems salinity is not directly measured but determined through a conductivity measurement. Even the term CTD is inaccurate, since depth is a distance, and a CTD does not measure its distance from the sea surface but employs a pressure measurement to indicate depth. But the three most important oceanographic parameters which form the basis of a regional description of the ocean are temperature, salinity, and pressure, which the CTD delivers.

In this text we follow oceanographic convention and express temperature T and potential temperature Θ in degrees Celsius (°C) and pressure p in kiloPascal (kPa, 10 kPa = 1 dbar, 0.1 kPa = 1 mbar; for most applications, pressure is proportional to depth, with 10 kPa equivalent to 1 m). Salinity S is taken to be evaluated on the Practical Salinity Scale (even when data are taken from the older literature) and therefore carries no units. Density ρ is expressed in kg m^{-3} or represented by $\sigma_t = \rho - 1000$. As is common oceanographic practice, σ_t does not carry units (although strictly speaking it should be expressed in kg m^{-3} as well). Readers not familiar with these concepts should consult textbooks such as Pickard and Emery (1990), Pond and Pickard (1983), or Gill (1982); the last two include information on the Practical Salinity Scale and the Equation of State of Seawater which gives density as a function of temperature, salinity, and pressure. We use z for depth (z being the vertical coordinate in a Cartesian xyz coordinate system with x pointing east and y pointing north) and count z positive downward from the undisturbed sea surface $z = 0$.

A CTD typically returns temperature to 0.003°C, salinity to 0.003 parts per thousand, and depth to an accuracy of 1 - 2 m. Depth resolution can be much better, and advanced CTD systems, which produce data triplets at rates of 20 Hz or more and apply data averaging, give very accurate pictures of the structure of the ocean along a vertical line. The basic CTD data set, called a CTD station or cast, consists of continuous profiles of temperature and salinity against depth (Figure 2.1 shows an example). The task of an oceanographic cruise for the purpose of regional oceanography is to obtain sufficient CTD stations over the region of interest to enable the researcher to develop a three-dimensional picture of these parameters and their variations in time. As we shall see later, such a data set generally gives a useful picture of the velocity field as well.

For a description of the world ocean it is necessary to combine observations from many such cruises, which is only possible if all oceanographic institutions calibrate their instrumentation against the same standard. The electrical sensors employed in CTD systems do not have the long-term stability required for this task and have to be routinely calibrated against measurements obtained with precision reversing thermometers and with salinometers, which compare water samples from CTD stations with a seawater standard of known salinity (for details see Dietrich *et al*., 1980). A CTD is therefore usually housed

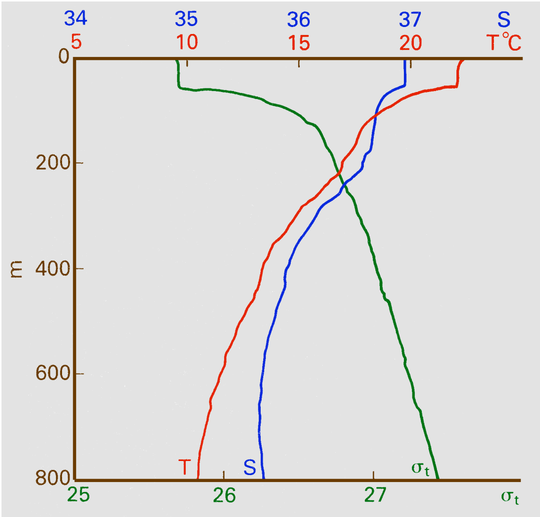


Fig. 2.1.

An example of the basic CTD data set. Temperature T and salinity S are shown against pressure converted to depth. Also shown is the derived quantity σ_t.

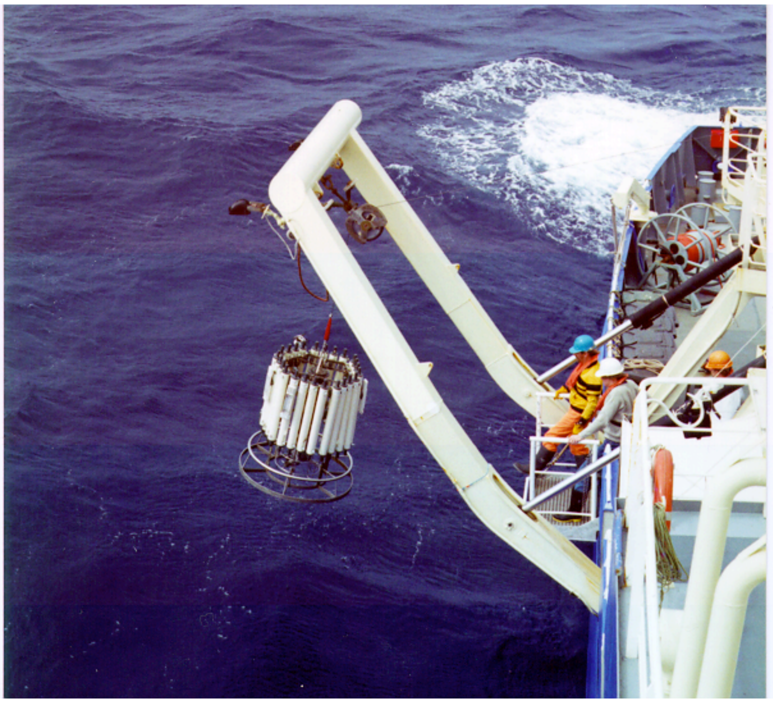


Fig. 2.2. A CTD is retrieved after completion of a station. The instrument is mounted in the lower centre, protected by a metal cage to prevent damage in rough weather. Above the CTD are 24 sampling bottles for the collection of water samples. The white plastic frames attached to some of them carry precision reversing thermometers.

inside a frame, with 12 or more bottles around it (Figure 2.2). The water samples collected in the bottles are used for calibration of the CTD sensors. In addition, oxygen and nutrient content of the water can be determined from the samples in the vessel's laboratory.

The CTD developed from a prototype built in Australia in the 1950s and has been a major tool of oceanography at the large research institutions since the 1970s. Two decades are not enough to explore the world ocean fully, and regional oceanography still has to rely on much information gathered through bottle casts, which produce 12 - 24 samples over the entire observation depth and therefore are of much lower vertical resolution. Although bottle data have been collected for nearly 100 years now, significant data gaps still exist, as is evident from the distribution of oceanographic stations shown in Figure 2.3. In the deep basins of the oceans, where variations of temperature and salinity are small, very high data accuracy is required to allow integration of data from different cruises into a single data set. Many cruise data which are quite adequate for an oceanographic study of regional importance turn out to be inadequate for inclusion in a world data set.

To close existing gaps and monitor long-term changes in regions of adequate data coverage, a major experiment, planned for the decade 1990 - 2000, is under way. This World Ocean Circulation Experiment (WOCE) will cover the world ocean with a network of CTD stations, extending from the surface to the ocean floor and including chemical measurements. Figure 2.4 shows the planned global network of cruise tracks along which CTD stations will be made at intervals of 30 nautical miles (half a degree of latitude, or about 55 km). As a result, we can expect to have a very accurate global picture of the distribution of the major oceanographic parameters by the turn of the century.

Because of the need for a global description of the oceanic parameter fields, researchers have attempted to extract whatever information they can from the existing data base.

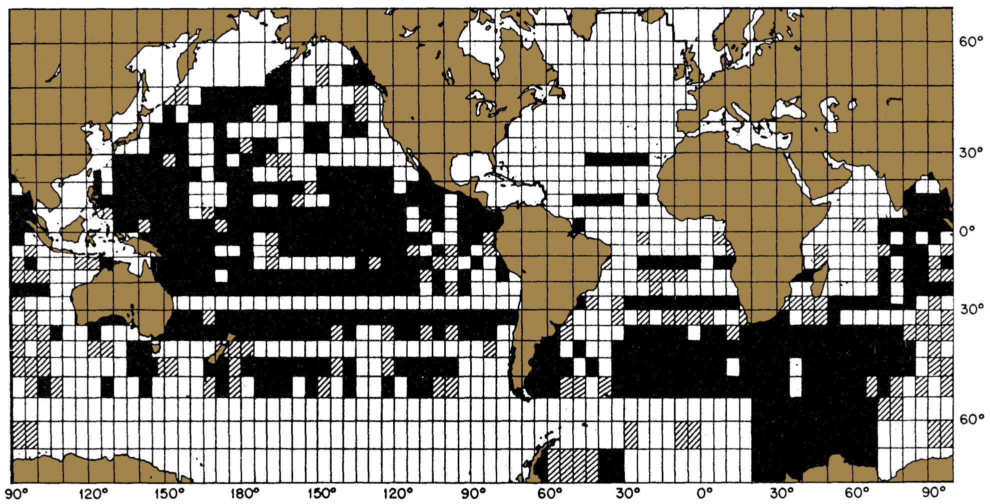


Fig. 2.3. World wide distribution of oceanographic stations of high data quality shortly before 1980. Unshaded 5° squares contain at least one high-quality deep station. Shaded 5° squares contain at least one high-quality station in a shallow area. Black 5° squares contain no high-quality station. Adapted from Worthington (1981)

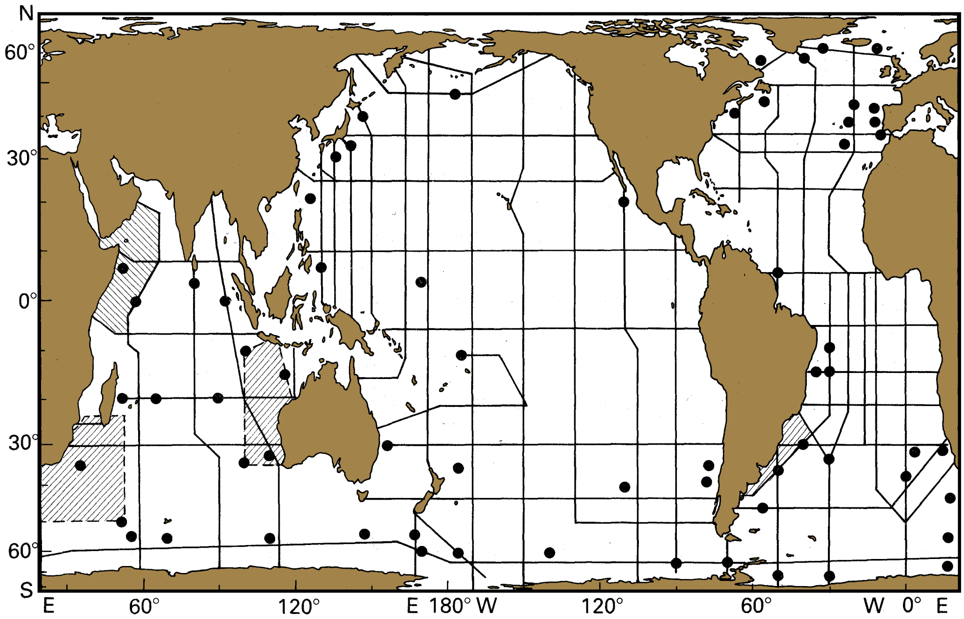
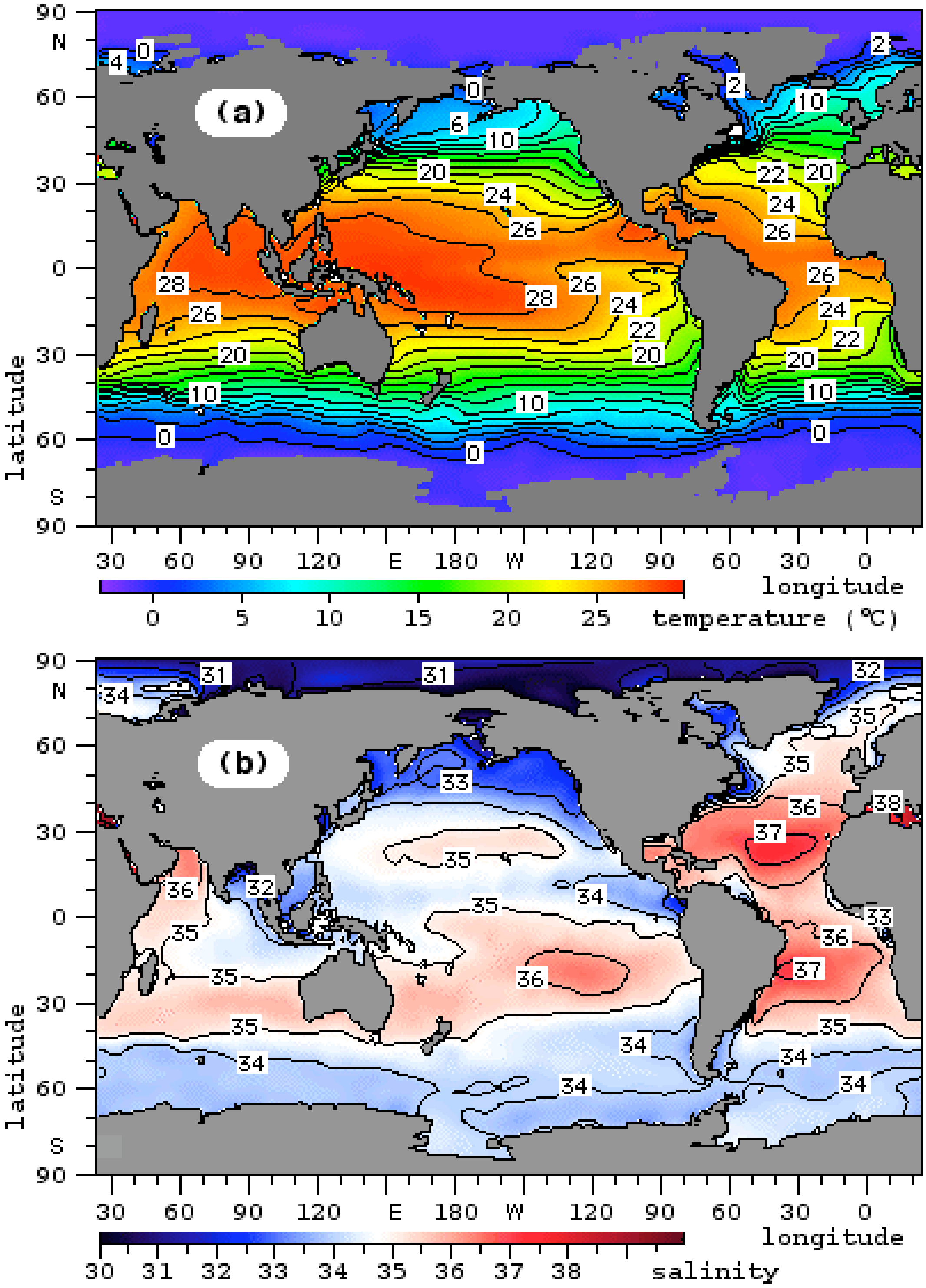


Fig. 2.4. The hydrographic sections of the World Ocean Circulation Experiment (WOCE). Shaded regions indicate intensive study areas. Dots indicate positions of current meter moorings.

Figure 2.5 is an example of a recent and widely used attempt. It includes all available oceanographic data regardless of absolute accuracy and shows that many features of the oceans can be studied without the very high data accuracy required for the analysis of the deep basins. Features such as the large pool of very warm surface water in the equatorial western Pacific and eastern Indian Oceans, the outflow of high salinity water from the Eurafrican Mediterranean Sea into the Atlantic Ocean below 1000 m depth, the formation of cold bottom water in the Weddell and Ross Seas near Antarctica, or the outflow of low salinity water from the Indonesian seas into the Indian Ocean, are all clearly visible in the existing data base. However, it should be remembered that the number of observations available for every 2° square varies considerably over the area and decreases quickly with

Fig. 2.5 (pages 19 – 21). Climatological mean potential temperature Θ (°C) and salinity S for the world ocean. Page 19: (a) Θ at $z = 0$ m, (b) S at $z = 0$ m, Page 20: (c) Θ at $z = 500$ m, (d) S at $z = 500$ m, Page 21: (e) Θ at $z = 2000$ m, (f) S at $z = 2000$ m. From Levitus (1982). The maps were constructed from mean values calculated from all available data for "2° squares", elements of 2° longitude by 2° latitude, and smoothed over an area of approximately 700 km diameter. Temperatures below 1°C are not plotted. The lowest values reached at the 2000 m level are around 0.0°C in the Antarctic and near -0.9°C in the Arctic region.


(a)
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S 90
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longitude
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temperature (°C)
(b)
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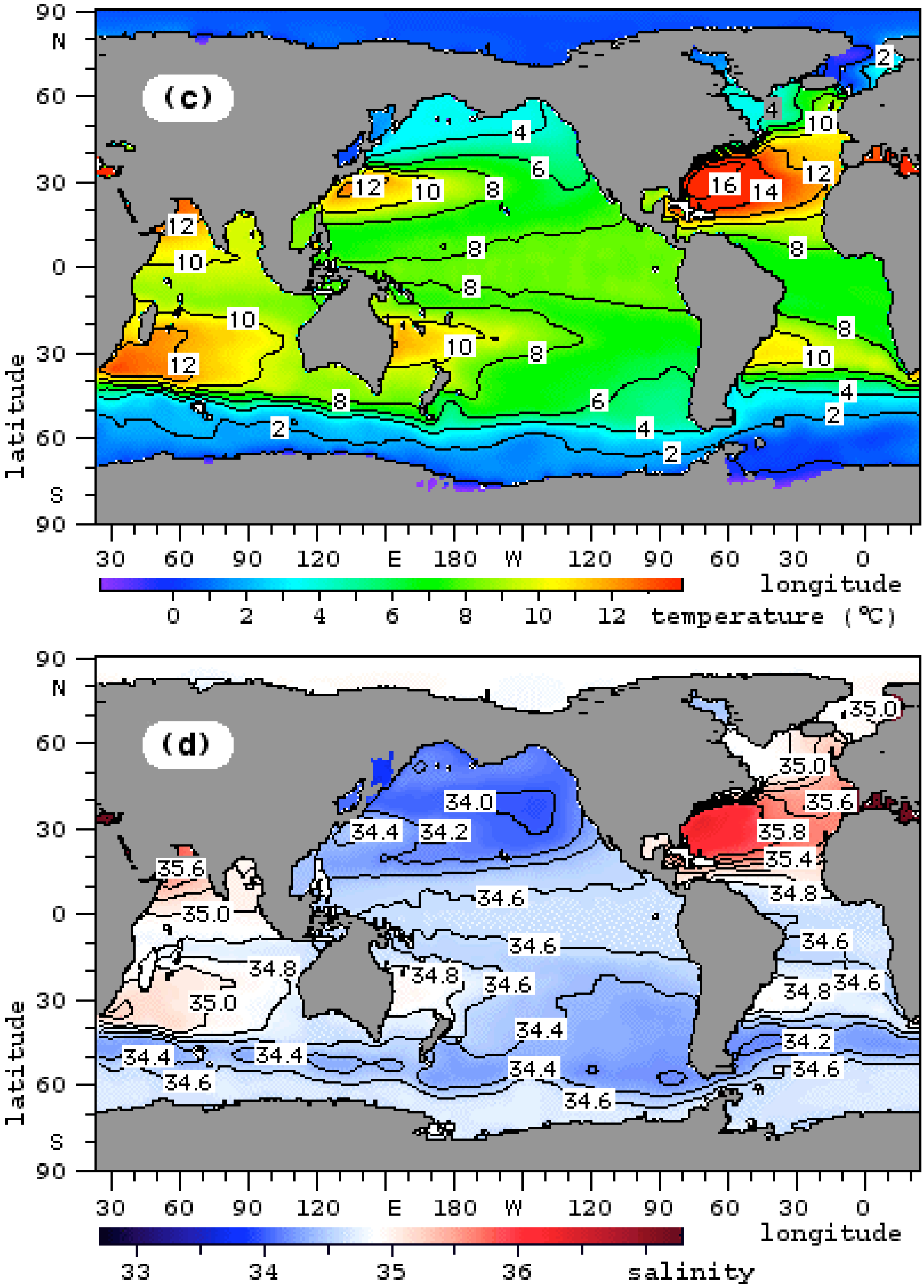
longitude
30 31 32 33 34 35 36 37 38
salinity


(c)
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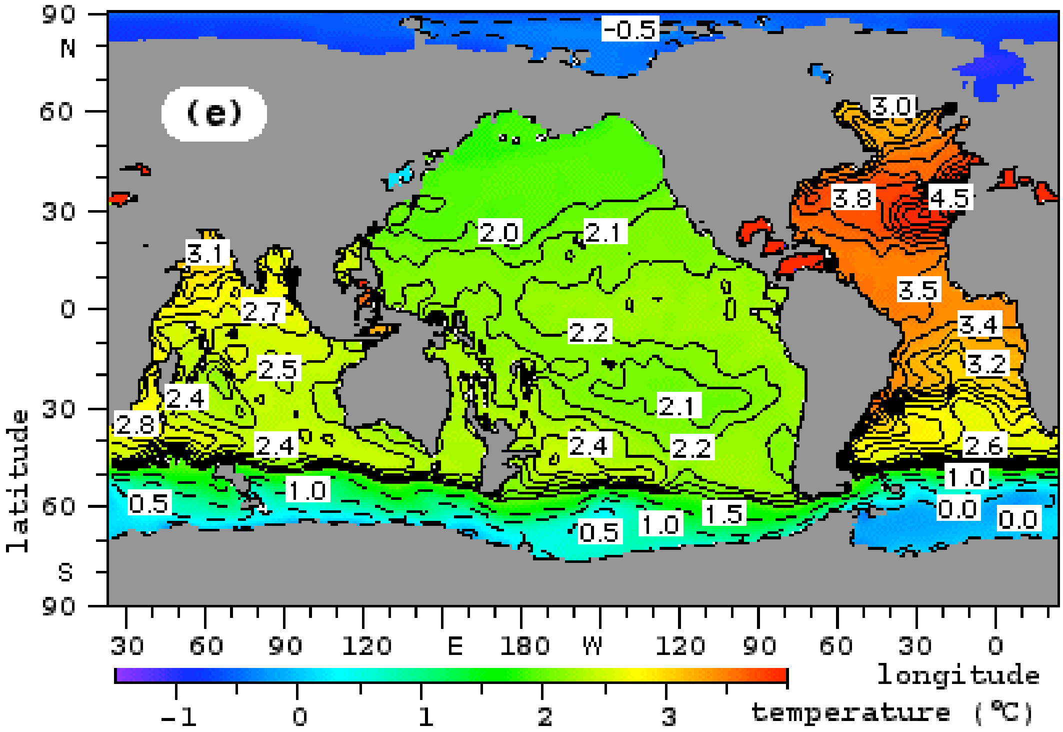
(e)
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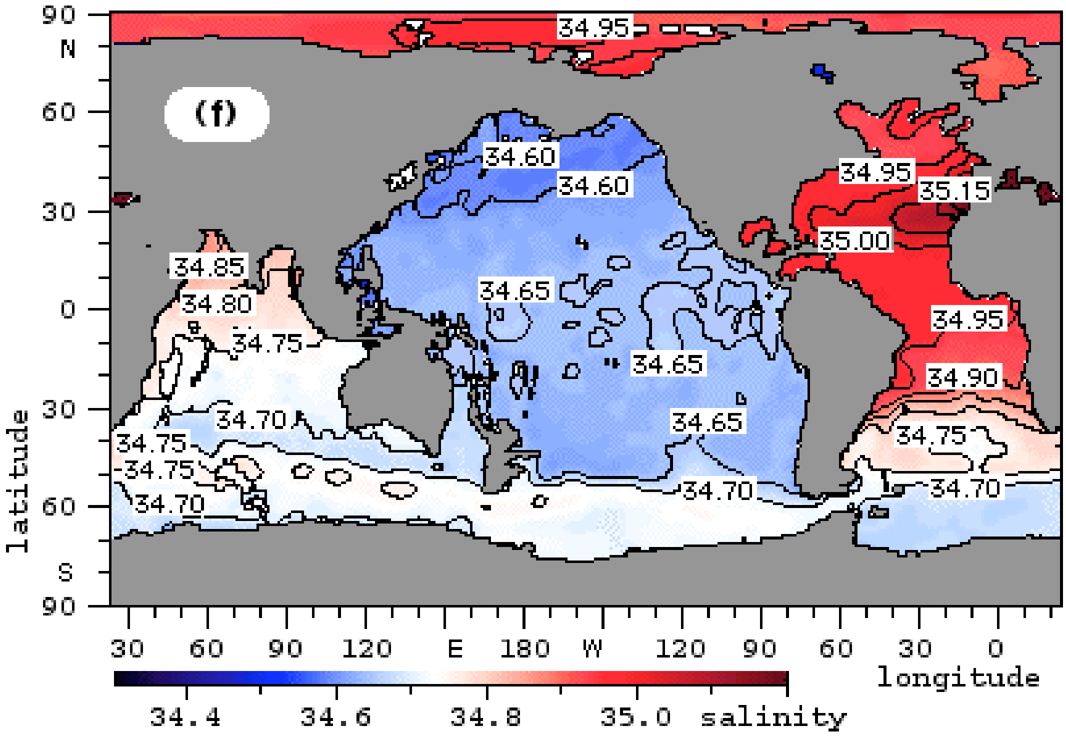
(f)
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34.70
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34.70
latitude
longitude
salinity

depth; in the polar regions it is also biased towards summer observations. Detailed interpretation of these and similar maps always has to take into account the actual data distribution.

Of itself, such information is only mildly interesting; but it is surprising what can be deduced from it. These deductions go into much more detail and reach much further than the examples just listed, which follow from simple qualitative arguments about the shape of isotherms or isohalines. More detailed analysis is based on the fact that most ocean currents can be adequately described if the oceanic pressure field is known (just as the atmospheric wind field follows from the air pressure distribution). Pressure at a point in the ocean is determined by the weight of the water above, which depends on the depth of the point and on the density of the water above it. As already noted, seawater density is a function of temperature, salinity and pressure. It is therefore possible - subject to some assumptions - to deduce the pressure distribution in the ocean and thus the current field from observations of temperature and salinity. How this is done is reviewed in the remainder of this chapter.

The first step in an accurate calculation of the oceanic pressure field is the calculation of density ρ from the Equation of State

$$\rho = \rho\ (\ T, S, p\ )\ . \qquad (2.1)$$

Much care has gone into the laboratory measurements of density as a function of tempera–ture T, salinity S and pressure p, and the Equation of State of Seawater now allows the calculation of density to a fractional accuracy of $3{\cdot}10^{-5}$ (0.03 kg m^{-3}) (Unesco, 1981; Millero and Poisson, 1981). We are now able to construct the density field to an accuracy comparable with the best field measurements of T, S and depth (or pressure p). The lowest accuracy is actually in the determination of depth since the pressure sensor is usually accurate to within 0.5 - 1% of full range, i.e. to 5 - 10 m if the sensor range is 1000 m. However, the oceanic pressure field can be determined with much higher accuracy from the distribution of density, as will be seen in a moment.

The quality of T and S measurements and of the presently-used equation of state can be checked by examining data collected in the high pressures found at great depth, where our measurement techniques are given their severest test. Figure 2.6 shows measurements from the vicinity of the deepest known place in the ocean. The measured temperature T increases with depth over the last six kilometers, while salinity S varies little. In a constant pressure environment this would indicate an apparent static instability, i.e. errors in the Equation of State. However, the same laboratory experiments that gave us the Equation of State allow us to determine very accurately the temperature drop that would occur if a parcel of water were brought to the surface without heat exchange. It would cool on decompression, depending on its salinity by 0.03 – 0.12°C per 1 km, and attain its "potential temperature" Θ. Figure 2.6 shows both T and Θ. It is seen that Θ is constant within measurement error. The "potential density" (the density the water would have if it were brought to the surface without changing salinity and potential temperature) is thus constant within measurement error, too. The available oceanic observations of today show that it is extremely rare to find inversions of potential density, i.e. situations where denser water appears to be lying on top of lighter water. Because in reality such inversions are unstable and overturn very quickly, their absence in observational data implies that eqn (2.1) obtained from laboratory measurements is in fact valid in the ocean.

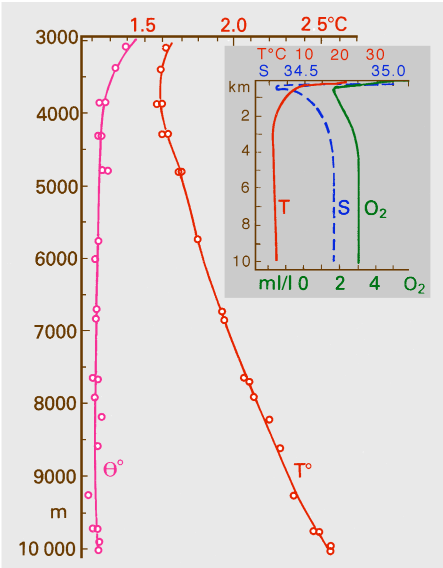

Fig. 2.6. Temperature T and potential temperature Θ in the Philippines Trench.

The inset also shows salinity S and oxygen O_2. From Bruun *et al.* (1956)

Precise knowledge of the density field is the basis for the second step, accurate calculation of the pressure field $p(z)$ from the hydrostatic relation

$$\partial p / \partial z = g \rho \quad , \qquad (2.2)$$

where g is gravity, g = 9.8 m s^{-2}, and depth z increases downwards. This equation is not uniformly valid (it does not hold for wind waves, for example); but it can be shown that it holds very accurately, to the accuracy of eqn (2.1), if it is applied to situations of sufficiently large space and time scales. It forms the basis of regional oceanography.

Evaluation of the pressure field from the hydrostatic equation involves a vertical integration of density. The advantage of an integration is that it eliminates the uncertainty in the measurement of depth as a source for inaccuracy. Its disadvantage is that it requires a reference pressure as a starting point. Without that information, eqn (2.2) can be used to get *differences* between pressures at different depths. An alternative way, which is common practice in oceanography, is to determine the distance, or depth difference, between two surfaces of constant pressure. For this purpose, a quantity called steric height h is introduced and defined as

$$h\,(z_1, z_2) = \int_{z_1}^{z_2} \delta\,(T,\,S,\,p)\,\rho_o dz \quad , \qquad (2.3)$$

where ρ_o is a reference density, and

$$\delta\ (T,\ S,\ p) = \rho\ (T,\ S\ p)^{-1} - \rho\ (0,\ 35,\ p)^{-1} \qquad (2.4)$$

is called the specific volume anomaly; it is the difference in volume between a unit mass of water at temperature T and salinity S and a unit mass at the standard salinity $S = 35.0$ and temperature $T = 0°C$. Steric height h has the dimension of height and is expressed in meters. To sufficient approximation,

$$\delta\ (T,\ S,\ p) = \frac{\rho_o - \rho\ (T,\ S,\ p)}{\rho_o^{\ 2}} \qquad (2.4a)$$

so that eqn (2.3) can also be written

$$h\ (z_1\ ,\ z_2) = \int_{z_1}^{z_2} \Delta\rho\ (T,\ S,\ p)\ /\ \rho_o(p)\ dz \quad , \qquad (2.3a)$$

where $\rho_o\ (p) = \rho\ (0,\ 35,\ p)$ and $\Delta\rho\ (T,\ S,\ p) = \rho_o - \rho\ (T,\ S,\ p)$. $h(\ z_1\ ,\ z_2)$ measures the height by which a column of water between depths z_1 and z_2 with standard temperature $T = 0°C$ and salinity $S = 35.0$ expands if its temperature and salinity are changed to the observed values. Typically, h is a few tens of centimeters. Because the weight of the water is not changed during expansion, the pressure difference between top and bottom remains the same. It is seen then that $h(z_1,z_2)$ measures variations in the vertical distance between two surfaces of constant pressure.

For oceanographic purposes, the sea surface can always be regarded as an isobaric surface. As any meteorological air pressure map (such as Figure 1.3) tells us, this is only an approximation. In reality, pressure differences between atmospheric highs and lows are of the order of 2 - 3 kPa. However, the ocean reacts to these differences by expanding in regions of low atmospheric pressure and contracting under atmospheric highs. These vertical movements are of the order of 0.2 - 0.3 m; like the tides, they have little effect on the long-term flow field and can be disregarded in our discussion, which is concerned with water movement induced by the oceanic pressure field.

Unfortunately, the sea surface cannot be used as the reference surface for the integration of eqns (2.3) or (2.3a) because it is not necessarily flat, and its exact shape is not known. We may, however, assume for the moment that a constant pressure surface which does not vary in depth can be found somewhere in the ocean and call it, for reasons which will become clear in a moment, a "depth of no motion". It then becomes possible using eqn (2.3) to map the oceanic pressure distribution by measuring, for every isobaric surface, its distance from the position z_0 of this depth of no motion. A possible pressure distribution is sketched in Figure 2.7. Because the weight of the water above the depth of no motion has to be the same everywhere, the sea surface, which is given by the steric height $h(0,\ z_o)$, is lower above the region of high water density than above the region with low water density. It is seen that a two-dimensional representation of the situation on a horizontal map is possible in two ways. We can either select a constant depth surface $z = z_r$ and map the intersections of the isobars with the depth surface, or we can select a constant pressure surface p_1 and draw contours of constant steric height. The first method is well known from meteorology; daily weather forecasts are based on maps of isobars at sea level (considered flat for the purpose of meteorology). In oceanography the position of the sea surface is unknown and has to be determined by analysis. Oceanographers therefore map the shape of the sea surface by showing contours of equal steric height relative to a depth of no motion, where pressure is assumed to be constant.

It is easy to show - subject to our assumption of a depth of no motion - that at any depth level, contours of steric height coincide with contours of constant pressure. The hydrostatic relation (eqn 2.2) tells us that if pressure is constant at $z = z_o$, the quantity $\rho_o\, g\, h\, (z, z_o)$ measures the pressure variations along a surface of constant height z. Thus, a contour map of h is an isobar map scaled by the factor $\rho_o\, g$ (see Figure 2.7).

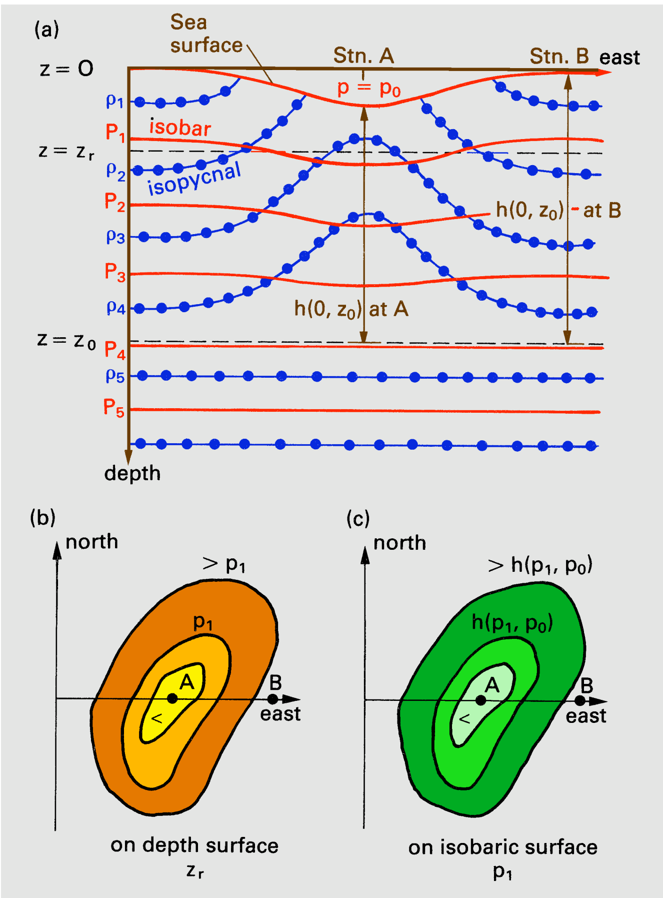

Fig. 2.7. Schematic illustration of steric height as a measure of distance between isobaric surfaces, and of the relationship between maps of isobars at constant height and maps of steric height at constant pressure.
(a) Distribution of isobars and isopycnals: at any depth level above $z = z_o$, water at station *A* is denser than water at station *B*. As the weight of the water above $z = z_o$ is the same, the water column must be longer at *B* than at *A*. The steric height of the sea surface relative to $z = z_o$ is given by $h(p_o, p_4)$, which in oceanographic applications is often given as $h(0, z_o)$, i.e. with reference to depth rather than pressure. The difference is negligible.
(b) The corresponding pressure map at constant depth $z = z_r$. (c) The corresponding map of steric height at constant pressure $p = p_1$.
The diagram requires some study, but it is well worth it; understanding these principles is the basis for the interpretation of many features found in the oceanic circulation.

A quantity widely found in oceanographic literature is the dynamic height D. It is defined as

$$D\,(p_1,\; p_2) = \int_{p_1}^{p_2} \delta\;(T,\,S,\,p)\;dp \qquad (2.5)$$

and is equal to gh, i.e. the product of gravity and steric height. Maps of dynamic height, also known as dynamic topography maps, are therefore maps of steric height scaled by the factor g or pressure maps scaled by the factor ρ_o. We prefer the use of steric height because it has the unit of length and therefore can be directly interpreted in terms of, for example, the shape of the sea surface. Other representations do, of course, just as well, as long as we remember that the "dynamic metre" often given as the unit for D is not a length but corresponds to $m^2\ s^{-2}$.

Is it possible to find a flat pressure surface in the ocean, i.e. one where the horizontal pressure gradient vanishes? One consequence of zero horizontal pressure gradient would be the absence of a current at that depth - hence the name "depth of no motion". Observations support the idea that in the deep ocean flow might, indeed, be so slow that the deep pressure map can be treated as flat. They show that below about 1300 m temperature and salinity are rather uniform, at least within a given basin. This comes out clearly in the maps of Figure 2.5, which would not show any structure at 2000 m depth outside the Southern Ocean if the relatively coarse contour interval of the 500 m maps were applied here. Furthermore the T and S gradients contribute roughly equal and opposite amounts to the density, so that density is remarkably uniform at such depths (even in the Southern Ocean). Within a basin, the density field is so horizontally uniform that steric height at 1500 m relative to 2000 m, which is shown in Figure 2.8a, displays horizontal variations of only a centimeter or so, within a given basin - and those variations look so random that they can be just as much a result of noise in the small data base as a real effect. By contrast, the steric height map for the sea surface relative to 2000 m (Figure 2.8b) shows differences of 0.5 m in a single basin and 1.8 m or so from highest to lowest point in the ocean, because the horizontal gradients of density are much greater (by a factor of several hundred) near the surface than at depths of 1500 - 2000 m.

These facts do not prove that the ocean is moving relatively slowly at depths of 1500 m or 2000 m; all they show is that if flow is slow at the one depth, it is slow at the other. However, in most parts of the ocean similar remarks apply for all pairs of depths (z_1, z_2) when both lie at or below about 1500 m, so the observations show that if there is any strong motion at these depths, it must take the form of a nearly vertically uniform flow everywhere below 1500 m. Because of the ocean's rough bottom it seems unlikely that such motions can be very strong or extend over great distances. Unfortunately, direct measurements of deep ocean currents are much harder to make than measurements of temperature and salinity; but in most parts of the ocean, such measurements as have been made support the idea that flow at these depths is very slow.

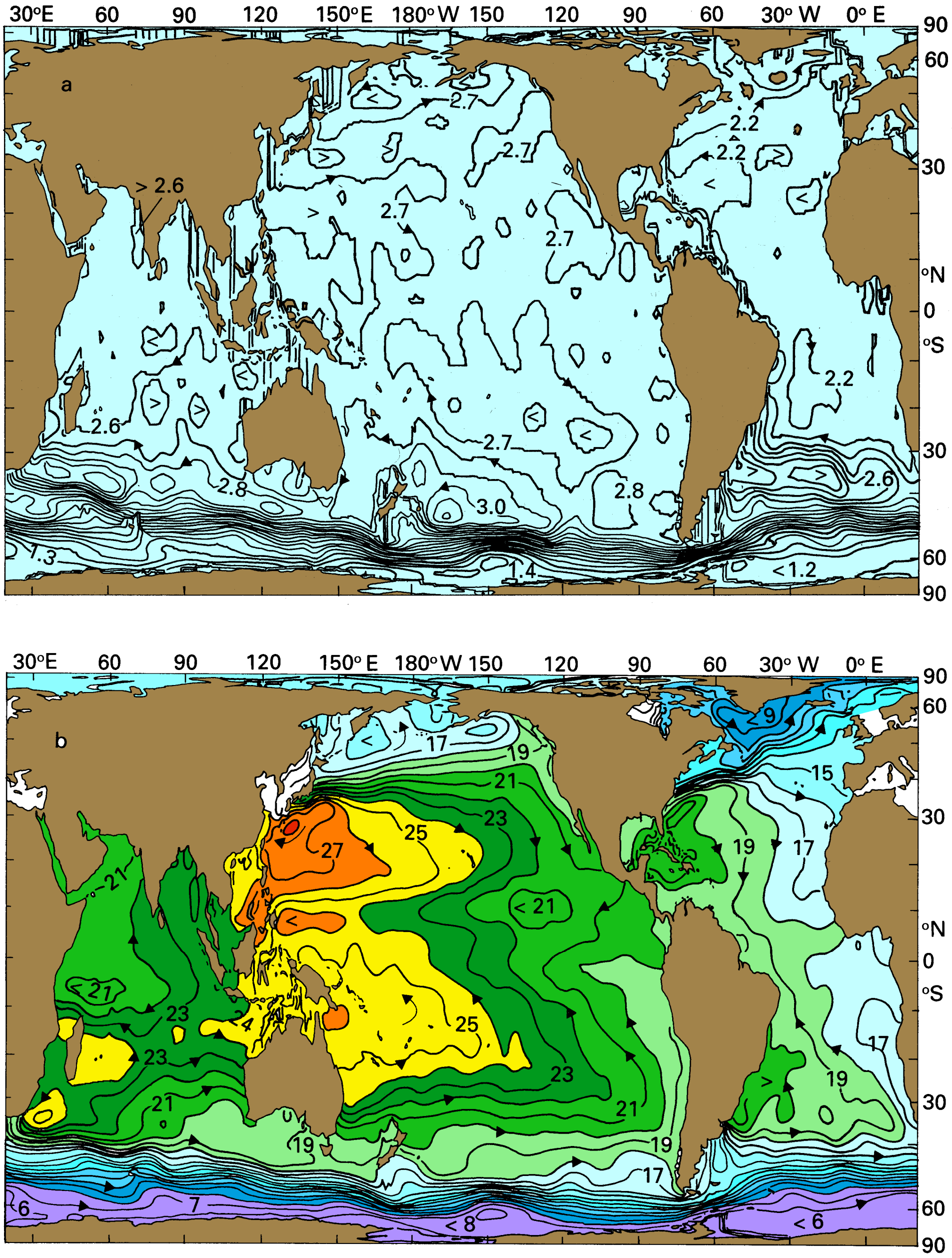

Fig. 2.8. Dynamic height ($m^2\ s^{-2}$), or steric height multiplied by gravity, for the world ocean. (a) at 1500 m relative to 2000 m, (b) at 0 m relative to 2000 m. Arrows indicate the direction of the implied movement of water, as explained in Chapter 3. (Divide contour values by 10 to obtain approximate steric height in m.) From Levitus (1982).