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PART I

THE OCEAN ENVIRONMENT

The following three chapters describe the average state of the ocean, its variability on a week-to-week and month-to-month basis (planetary waves and eddies), and its short-term variability (internal waves). The subsequent applications to ocean acoustics refer specifically only to the beginning of § 1.1, and §§ 1.6 and 3.6. This is because we place our emphasis (rather too heavily) on the role played by internal waves, and rely on a special model spectrum which illustrates our results by specific examples. Further, internal waves *are important*, as indicated in Part V by some striking agreements between measured and computed sound fluctuations.

But there is more to the ocean than internal waves! We have attempted in the first two chapters to give a description of the real ocean variability against which the application of a universal internal-wave spectrum can be judged, and to provide the background for future work involving the important effects of microstructure, ocean fronts, the surface mixed layer, and mesoscale variability. We make no pretense at completeness; this would involve all of oceanography. We totally omit a discussion of the surface and bottom boundaries and their role (sometimes crucial) to sound transmission. All biological factors are ignored.

The material in the first two chapters is descriptive, including dynamics only when it helps to systematize observed facts.

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CHAPTER 1

OCEAN STRUCTURE

It used to be thought that the ocean, particularly the deep ocean, was uniform, but this was a case of mistaking lack of information for lack of features. In fact, the ocean waters are everywhere structured, with characteristic scales ranging from centimeters to the scale of ocean basins themselves. On the large scale the variability derives from an east–west asymmetry of the general circulation, and from distinct water masses produced in the Antarctic, Arctic, equatorial regions, and the marginal seas. But the production of mean-square gradients must be balanced on the average by an equivalent dissipation of mean-square gradients, and this inevitably leads to a cascade of scales into fine- and microstructure.

1.1. Scales

The oceans rotate and are stratified. The frequencies

$$\omega_i = 2\Omega \sin (\text{latitude}), \quad n = \left(-\frac{g}{\rho}\frac{d\rho}{dz}\right)^{\frac{1}{2}} \tag{1.1.1}$$

are convenient measures of rotation and stratification. The inertial (or Coriolis) frequency ω_i is twice the vertical component of the Earth's angular velocity Ω. Accordingly, $\omega_i u$ is the Coriolis 'force' associated with a unit mass moving horizontally with speed u.

We will call n the buoyancy frequency (it is more often called the Brunt–Väisälä frequency). The fluid is statically stable when n is real, that is when the potential density increases with depth $-z$. A balloon filled with water at some depth and then displaced vertically will oscillate at a frequency n (under idealized conditions).

The inertial frequency increases from zero at the equator to $2\Omega = 2\pi/(12 \text{ hours}) = 1.46 \times 10^{-4}\ \text{s}^{-1}$ at the poles, therefore the appropriate length scale is A, the radius of the Earth. The buoyancy

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frequency varies typically from 3 cph near the surface to 0.2 cph near the bottom; the fractional rate of decrease diminishes with depth, and can be roughly modeled by $n(z)=n_0\, e^{z/B}$, with B of order 1 km. Accordingly,

$$A=6370 \text{ km}, \quad B\approx 1 \text{ km} \tag{1.1.2}$$

are the length scales associated with rotation and stratification respectively. Typical values for the frequencies are (at 30° latitude and 1 km depth)

$$\omega_{\mathrm{i}}=\tfrac{1}{24} \text{ cph}, \quad n=1 \text{ cph} \tag{1.1.3}$$

so that in this sense the Earth is a slowly spinning planet; ω_{i}/n (typically 0.04) is a measure of the aspect ratio (vertical to horizontal) of ocean inhomogeneities; there is no justification for the usual assumption of spherical symmetry.

The four scales ω_{i}, n, A, B are fundamental to the following discussion of ocean fluctuations. Planetary (or Rossby) waves have frequencies below ω_{i}. The ocean circulation is dominated by mesoscale eddies (associated with planetary waves) that have correlation distances and correlation times of order

$$2\pi Bn/\omega_{\mathrm{i}}\approx 100 \text{ km}, \quad n^{-1}A/B=1000 \text{ hours}$$

respectively (Chapter 2). Internal waves have frequencies lying between ω_{i} and $n(z)$. These are discussed in detail in Chapter 3.

Sound speed and stratification

It is convenient to introduce the *potential* gradient; defined as the measured gradient minus the adiabatic gradient, the latter arising from the adiabatic expansion or compression of a rising or sinking volume. Accordingly, the vertical gradient in density can be written as a sum of potential and adiabatic gradients,

$$\partial_z\rho=\partial_z\rho_{\mathrm{P}}+\partial_z\rho_{\mathrm{A}},$$

with similar relations for temperature, T, and sound speed, C. Only the potential density gradient contributes to the stability of the water column. Similarly, only the potential gradient in sound speed, $\partial_z C_{\mathrm{P}}$, contributes to sound fluctuations associated with internal waves and other forms of vertical motion. Aside from these fluctuations it is the true sound speed $C(z)$ that determines the properties of the sound channel.

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Table 1.1. *Values of* $10^6 a$ *in* $\deg^{-1}$, *at stated temperatures and depths.* (*From Munk*, 1974.)

Depth (km)	Temperature (°C) 0	5	10	15	20	25
0.0	51	107	158	204	245	283
1.0	76	127	173	216	254	290
2.0	100	146	189	227	263	296
3.0	122	164	203	238	271	302
4.0	143	181	217	249	279	308
5.0	161	197	229	258	286	314

Potential density is a function of potential temperature and salinity, $\rho_P(T_P, S)$, and so

$$n^2(z) = -g\rho^{-1}\,\partial_z\rho_P = g(a\,\partial_z T_P - b\,\partial_z S) \tag{1.1.4}$$

where a and b are the coefficients of thermal expansion and saline contraction respectively. Similarly, the potential gradient in the fractional sound speed can be written

$$C^{-1}\,\partial_z C_P = \alpha\,\partial_z T_P + \beta\,\partial_z S. \tag{1.1.5}$$

Typical numerical values are

$$\begin{aligned} \alpha &= 3.19\times10^{-3}\,(°\mathrm{C})^{-1}, \quad \beta = 0.96\times10^{-3}\,(‰)^{-1} \\ a &\approx 0.13\times10^{-3}\,(°\mathrm{C})^{-1}, \quad b = 0.80\times10^{-3}\,(‰)^{-1}. \end{aligned} \tag{1.1.6}$$

The a-value is typical of conditions at 1 km depth, but varies considerably with temperature and pressure (Table 1.1).

The ratio

$$r = b\,\partial_z S/a\,\partial_z T_P \tag{1.1.7}$$

gives the relative contributions of salt and (potential) temperature to the stability of the water column.

We define

$$G = \frac{\alpha}{a}\frac{s(r)}{g}, \quad \frac{\alpha}{a} = 24.5, \quad s(r) = \frac{1+cr}{1-r}, \quad c = \frac{a\beta}{\alpha b} = 0.049, \tag{1.1.8}$$

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using the numerical values (1.1.6). Then

$$C^{-1}\,\partial_z C_{\mathrm{P}} = Gn^2(z). \qquad (1.1.9)$$

In any one location, G generally stays within relatively narrow limits, so that (1.1.9) expresses the potential sound-speed profile in terms of the stratification (which is perhaps the most fundamental aspect of ocean structure).

The total gradient in sound speed is the sum of the potential and adiabatic gradients,

$$\partial_z C = \partial_z C_{\mathrm{P}} + \partial_z C_{\mathrm{A}},$$

where

$$\begin{aligned} C^{-1}\,\partial_z C_{\mathrm{A}} &= \alpha\,\partial_z T_{\mathrm{A}} + \gamma\,\partial_z P \\ &= (-0.03 - 1.11)\times 10^{-2}\,\mathrm{km}^{-1} \\ &= -1.14\times 10^{-2}\,\mathrm{km}^{-1} \equiv -\gamma_{\mathrm{A}} \qquad (1.1.10) \end{aligned}$$

is the fractional sound-speed gradient in an adiabatic, isohaline ocean. The total sound speed therefore increases with increasing T, S and P. Unlike C_{P}, C has a minimum value $C = C_1$ at some depth $z_1 = -h$ and increases by a few per cent towards top and bottom. The increase above the minimum is the result of increasing temperature towards the surface; the increase below the minimum is associated with increasing pressure in the nearly isothermal deep waters.

Vertical fluxes

What are the physical processes responsible for the stratification? Fig. 1.1 shows the interior distribution (ignoring the messy surface layers) of potential temperature and salinity in the eastern North Pacific. From the T–S diagram (Fig. 1.2) we obtain $r \approx -0.55$, so that a fraction $-r/(1-r) = 0.35$ of the density stratification is due to salinity.† At abyssal depths conditions are fairly uniform, characterized by 1.1 °C potential temperature and 34.69 ‰ salinity. This is Antarctic Circumpolar Water spreading northward into the deep-basin of the world's oceans. The formation of new bottom water is accompanied by upwelling throughout the water column. We can

† At 40° S in the Pacific, $r \approx -0.44$. In the North Atlantic the salinity partially destabilizes; a typical value for Bermuda is $r = +0.8$.

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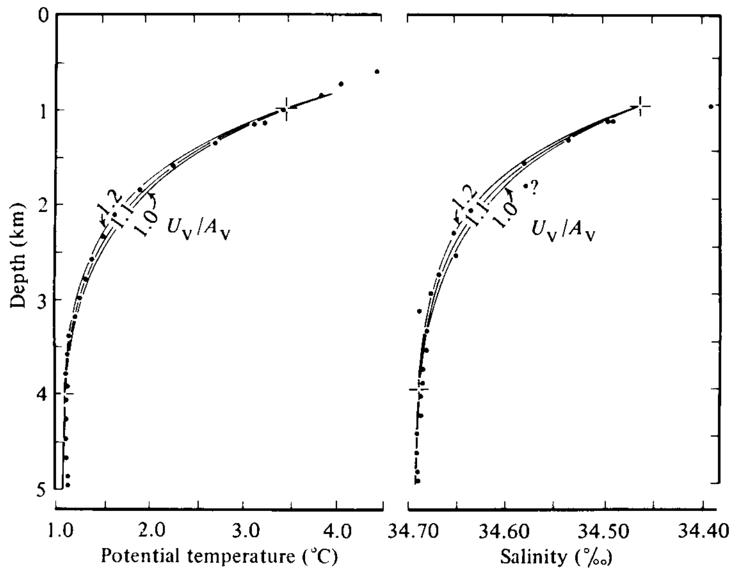

Fig. 1.1. Potential temperature and salinity as functions of depth at station CALCOFI 60–190, 33° 17′ N, 132° 42.5′ W (salinity at depth 1859 m was questioned in the original observations). Curves are labeled in reciprocal scale depth, units km^{-1}.

postulate (rather naively) that the observed vertical distribution represents a balance between vertical convection and diffusion:

$$[A_V\, d^2/dz^2 - U_V\, d/dz]\,T, S = 0, \qquad (1.1.11)$$

with temperature and salinity fixed at top and bottom. For constant vertical eddy diffusivity, A_V, and upwelling, U_V, this leads to an exponential distribution with depth scale $A_V/U_V \approx 1$ km (Fig. 1.1). In the case of nonconservative quantities (such as ^{14}C) we need to include the decay constant D (1.24×10^{-4} years^{-1}) as an additional term within the brackets of (1.1.11), leading to a time scale $A_V/U_V^2 \approx 200$ years (Munk, 1966). Hence, with A_V/U_V and A_V/U_V^2 given,

$$A_V \approx 1\ \text{cm}^2\,\text{s}^{-1}, \quad U_V \approx 1\ \text{cm day}^{-1} = 5\ \text{km}/(1000\ \text{years}). \qquad (1.1.12)$$

This model predicts a complete cycling of ocean water every one thousand years.

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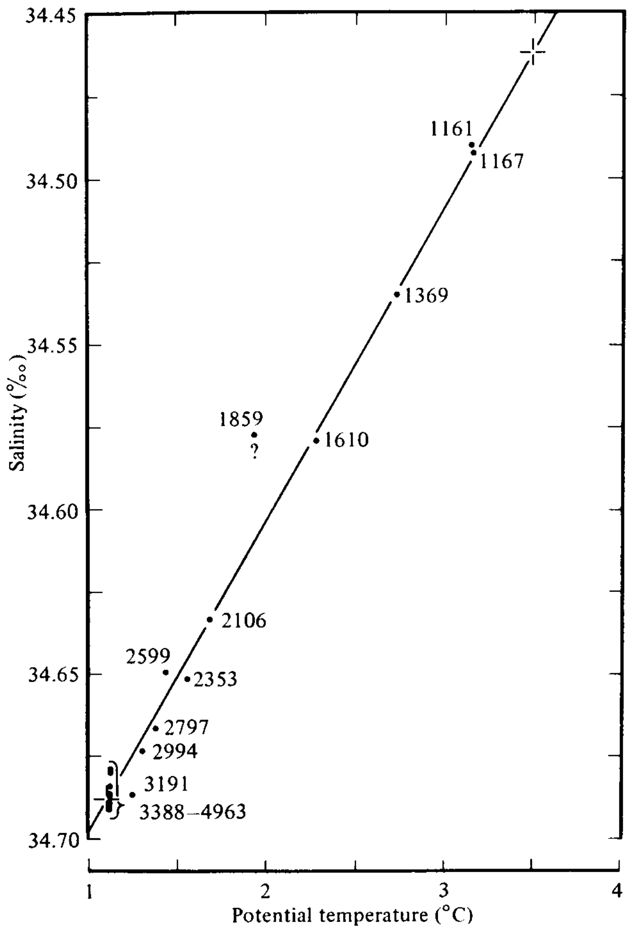

Fig. 1.2. The potential temperature versus salinity relation for station CALCOFI 60–190, with depth indicated in meters.

There are many things wrong with this picture of a vertical diffusive–convective model driven by the formation of Antarctic Bottom Water. Other water masses intrude at intermediary depths and modify the assumption of a constant U_V. The parameterization of vertical diffusion by a constant A_V is certainly inadequate, though the processes are not yet understood. (Leading theories

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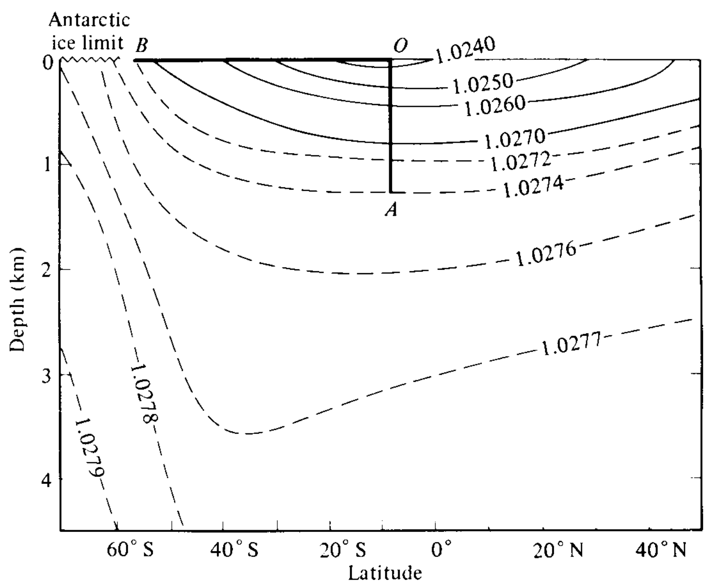

Fig. 1.3. Schematic representation of potential density in a north–south section of the Pacific.

include double diffusion discussed in Turner, 1973; and internal-wave breaking.) Finally, we have ignored all lateral processes.

Lateral fluxes

Referring to the schematic presentation of Fig. 1.3, we find that the T–S relation along the vertical section OA is very similar to the section OB along the sea surface. (This mapping of latitude into depth was noted by Iselin, 1939, in the Sargasso Sea.) The interpretation is that the water attains T–S characteristics at the air–sea boundary, and these are then readily communicated into the interior along surfaces of constant potential density. The profile of $\rho = 1.0274$ emerges just north of the antarctic ice limit, and we may expect the water above this level (about 1 km) to be strongly influenced by surface heating and cooling, precipitation and evaporation. This is discussed in the following section. There is no good agreement as to the relative importance of vertical and horizontal mixing processes.

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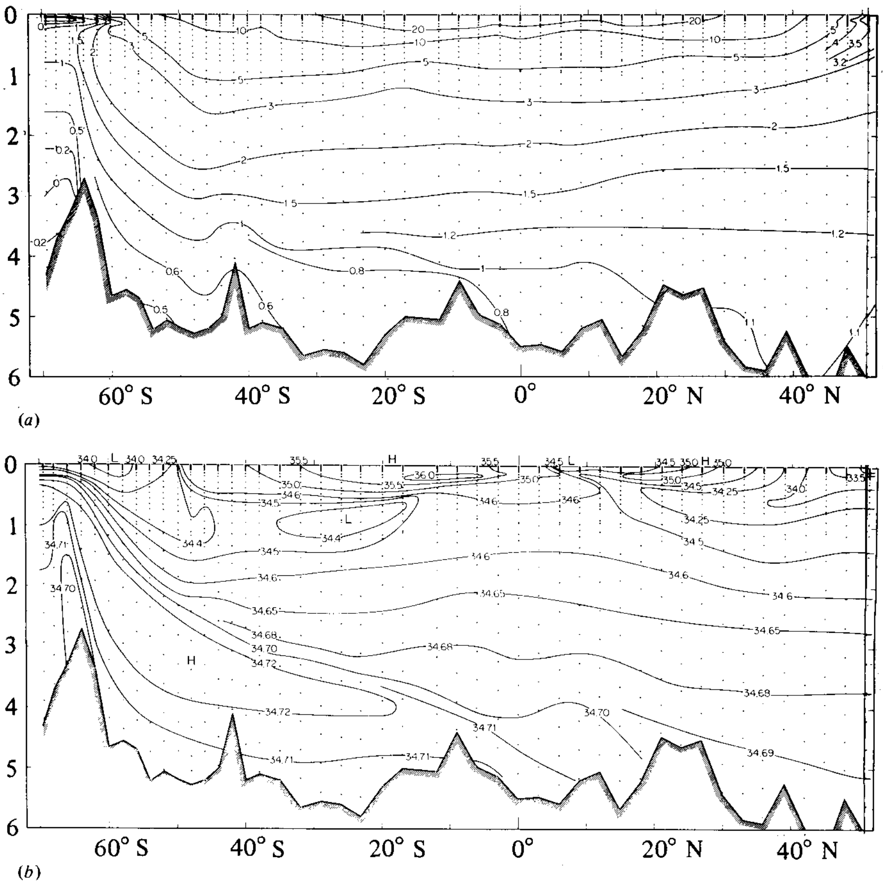

Fig. 1.4. Latitude–depth (in km) profiles in the Central Pacific (170° W), from Antarctica to the Aleutians. (*a*) Potential temperature contour in °C; (*b*) salinity contour in ‰; (*c*) sound-speed contour (from 1450 to

1.2. Water masses

Latitudinal variation of ocean structure is illustrated by an antarctic to arctic section through the Central Pacific (Fig. 1.4†). Observations north of 34° N were taken during the *Great Bear Expedition*, moving southward, April to June 1970 (Horibe, 1971); the remain-

† Joseph Reid has selected the data for the construction of these sections. We are grateful for his help and advice.

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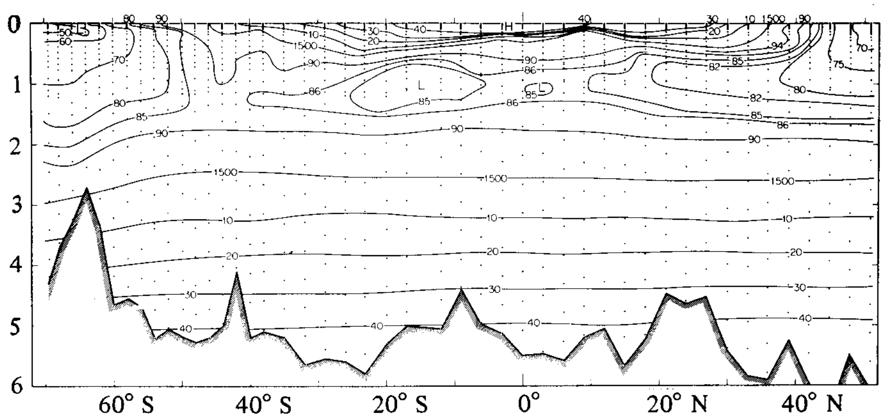
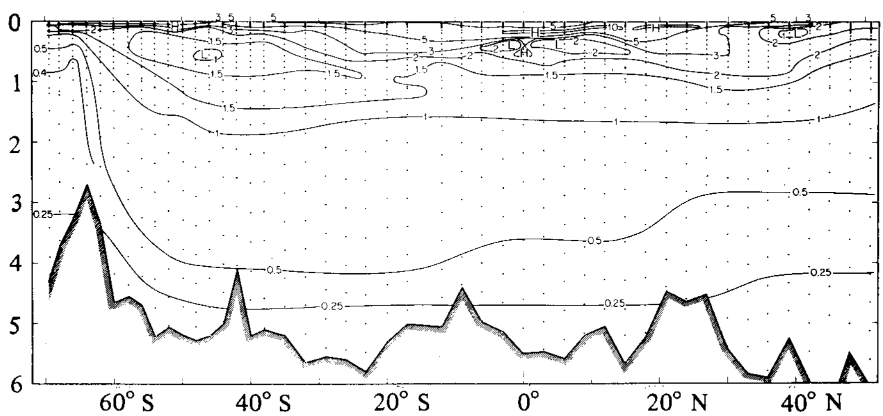


(c)



(d)

1540 m s^{-1}, the first two digits are omitted); and (*d*) buoyancy frequency contour in cph. Vertical exaggeration is about 1000 : 1; dots give measurement grid.

ing observations were taken during the *Southern Cross Cruise*, moving southward, November 1968 to March 1969 (Horibe, 1970).

Bottom water south of the Equator is characterized by about 0.6 °C, 34.71 ‰. This water is a derivative of the Antarctic Bottom Water which is formed mostly within the Weddell Sea by freezing processes (not yet completely understood) that generate a residual of salty, cold and therefore dense water (Gill, 1973). By the time it has moved (within the Antarctic Circumpolar Current) into the

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