

## CHAPTER 3 -PHYSICAL PROPERTIES OF SEAWATER

Historically ocean temperature and salinity profiles were measured by arrays of Nansen bottles, with reversing thermometers. A typical deep ocean cast would consist of the deployment of 20 to 30 bottles on a steel winch cable; from which temperatures with typical precisions of  $\pm 0.01$  °C and salinities with precisions of  $\pm 0.01$ psu could be obtained. The difference between protected and unprotected reversing thermometers was used to infer pressure (and hence depth) to within  $\pm 2$ m. Historically, chemical titration methods, assuming constant proportions among the different ions in seawater, were used to determine salinity according to the relation

$$S(\text{psu}) = 0.03 + 1.805 \text{Cl}(\text{‰}),$$

where Cl (parts per thousand) is the “chlorinity” of a water sample.

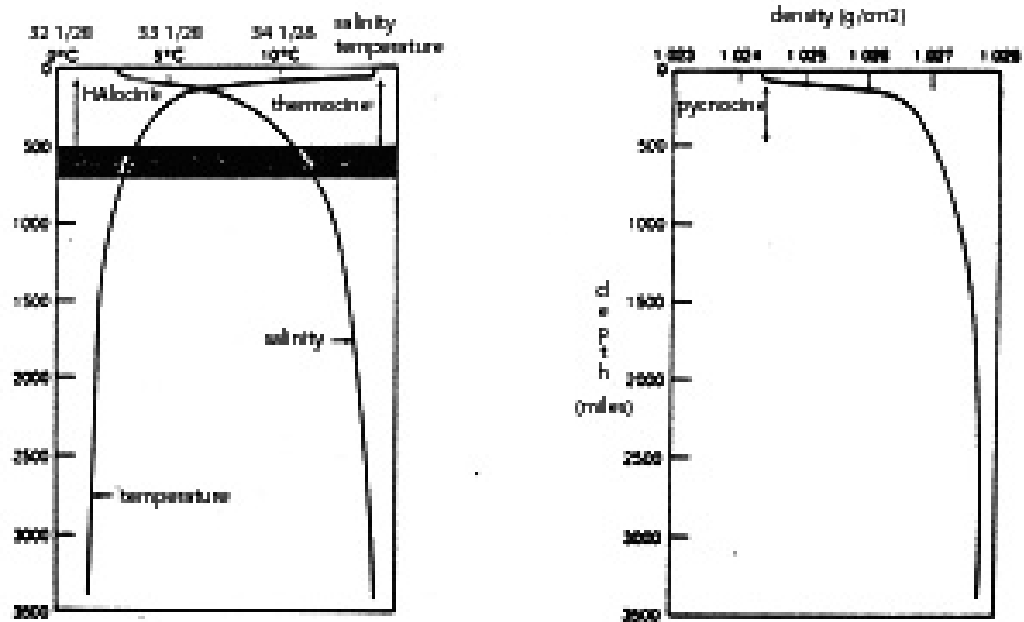
The modern oceanographic instrument used to measure continuous distributions of conductivity, temperature versus depth; (actually pressure) is called a CTD. The variety of electronic CTD instruments commercially-available today have enough resolution to resolve differences of 0.01 °C to 0.001°C in temperature ; equivalent 0.01 psu to 0.001 psu in salinity; and 0.10m to 1.00m in pressure on centimeter to meter-vertical scales.

Salinity is determined from precision measurements of seawater conductivity C and temperature T and well-known, complicated relationships that can be approximated for our purposes by:

$$S(\text{psu}) = 3.55 + 1.02 C \left( \frac{\text{mmho}}{\text{cm}} \right) - 0.73 T(\text{°C}).$$

Such measurements were used to infer the generic temperature, salinity and density profiles in [Figure 3.1](#). The upper ocean warmth, freshness and “lightness” relative to the deeper ocean are major features of these profiles. Later we will explore the real differences that exist throughout the world’s oceans.

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**Figure 3.1** Generic subtropical temperature, salinity, and density profiles exhibit regions of large change; thermocline, halocline, and pycnocline, respectively. (Duxbury and Duxbury, 1984)

### Seawater Density

Probably the most important dynamic property of seawater is its density  $\rho$ . In general the observed differences in seawater density are due to the combined effects of variability in temperature  $T$ , salinity  $S$ , and pressure  $P$ . Though relatively small (the range of sea surface densities is from 1.024 to 1.030 gm/cm<sup>3</sup>) horizontal differences in density produce horizontal pressure forces which in turn produce ocean currents.

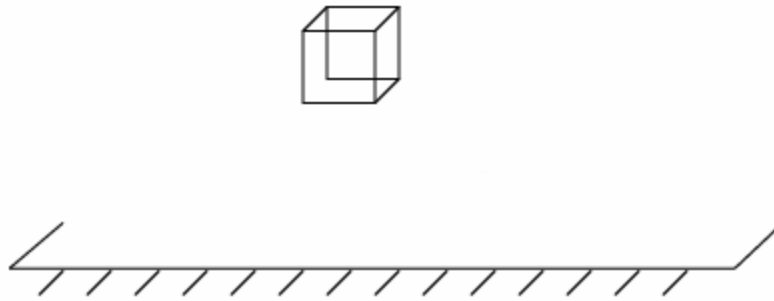
Symbolically the following relation

$$\rho = \rho(S, T, P) \text{ (or } \rho_{STP}\text{)}$$

expresses the dependence of the seawater density (mass per unit volume) of a water parcel (Figure 3.2) on its measured or *insitu* salinity,  $S$ , *insitu* temperature,  $T$ , and *insitu* pressure,  $p$ , where insitu refers to values of these variables at the location of the

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water parcel of interest.



**Figure 3.2** A parcel of seawater that is small enough so that the measured or *insitu* temperature T, salinity S, and pressure P is homogeneous for practical purposes.

Changes in the value of *insitu* density can be explored by considering its total differential,  $d\rho$ , as follows

$$d\rho = \left. \frac{d\rho}{dS} \right|_{T,P} dS + \left. \frac{d\rho}{dT} \right|_{S,P} dT + \left. \frac{d\rho}{dP} \right|_{S,T} dP.$$

Variability in density can be explored more conveniently in terms of *insitu* density anomaly,  $\sigma_{\text{STP}}$ , which is defined as

$$\sigma_{\text{STP}} = (\rho_{\text{STP}} - 1) \times 10^3.$$

Thus surface density anomalies vary from 24.0 to 30.0  $\text{mg}/\text{cm}^3$ .

### **Compressibility (or pressure) effects**

The most important (or first order) effect of hydrostatic pressure on density is to decrease the volume of a water parcel by squeezing the same molecules closer together;

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thus there is an increase in  $r_{STP}$ .

For example, if we consider what happens to a water parcel at the sea surface (with the given properties) after we move adiabatically (without heat loss or gain) to a depth of 4000 m where the pressure is ~400 atmospheres or 6000 psi (Figure 3.3).

In this example the pressure effect on  $s_{STP}$  is nearly 100%. In general this very large pressure effect obscures density changes due to the natural variability of temperature and salinity. Therefore oceanographers find it convenient to compare the density of water parcels in terms of the quantity called sigma-t and defined

$$s_T = s_{STP=0}$$

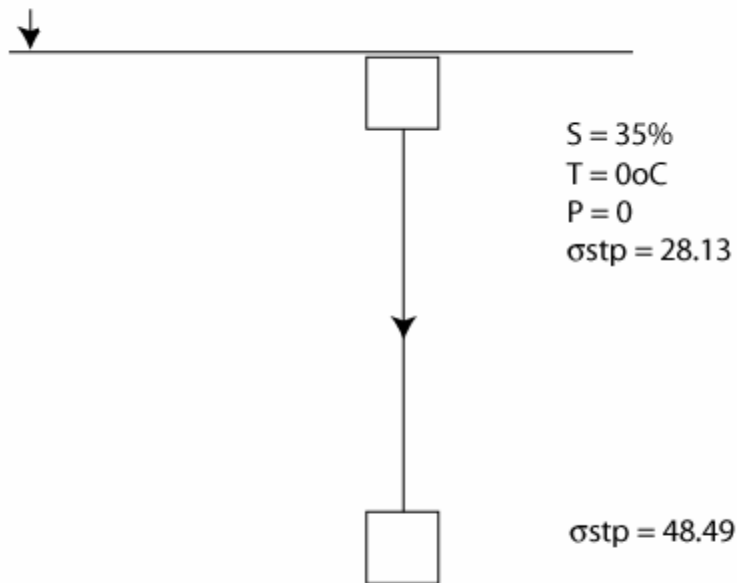
where P is gauge pressure which is zero at the sea surface.

This definition in effect removes the very large pressure effect on  $s_{STP}$  by evaluating  $s_{STP}$  at the *insitu* values of S and T and the sea surface pressure.  $s_T$  is a complicated function of S and T and can be found in Table 11-1, Knauss.

A secondary (or second order) effect of an increase in hydrostatic pressure is the increase of measured temperature,  $T_{insitu}$ , at the approximate rate of 0.1°C per kilometer of depth increase. The adiabatic compression of a water parcel causes increased molecular activity and thus an increased observed temperature.

For example, surface values of  $S_{surf} = 35\%$  and  $T_{surf} = 5.00^\circ\text{C}$  an adiabatic compression to a depth of 4000 m will produce a  $T_{insitu} = 5.45^\circ\text{C}$ . Conversely an adiabatic expansion from a depth of 4000 m with  $T_{insitu} = 5.00^\circ\text{C}$  will lead to  $T_{surf} = 4.56^\circ\text{C}$ . This non-linear adiabatic heating and cooling is reasonably well-understood and is summarized in Table 11-6, Knauss.

These tables can be used to compute a quantity called potential temperature,  $q$ , or  $T_{surf}$ . Potential temperature is  $T_{insitu}$  which has been corrected for compressibility effects.



**Figure 3.3.** The dependence of seawater density anomaly on pressure.

When  $q$  is used rather than  $T_{\text{insitu}}$  to compute density anomaly the quantity potential density anomaly, or sigma- $q$  or  $s_q$  is appropriate and is defined

$$s_q = s_{S,q,P=0}$$

Thus we have defined a density-related term,  $s_q$ , whose value depends on the intrinsic properties of a water parcel,  $q$  and  $S$ , (determined by air-sea interaction processes at the sea surface) only. The external influence of pressure on volume and temperature of the water parcel have been eliminated. The use of  $s_T$  and  $s_q$  is convenient when assessing the stability of a water column.

This comparison of  $T$  and  $q$  and  $s_T$  and  $s_q$  for data from the North Pacific in [Figure 3.4](#) shows the important the pressure effects on temperature, especially for

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depths greater than 4 km. Note the misleading stability implied by the  $s_T$  profile.

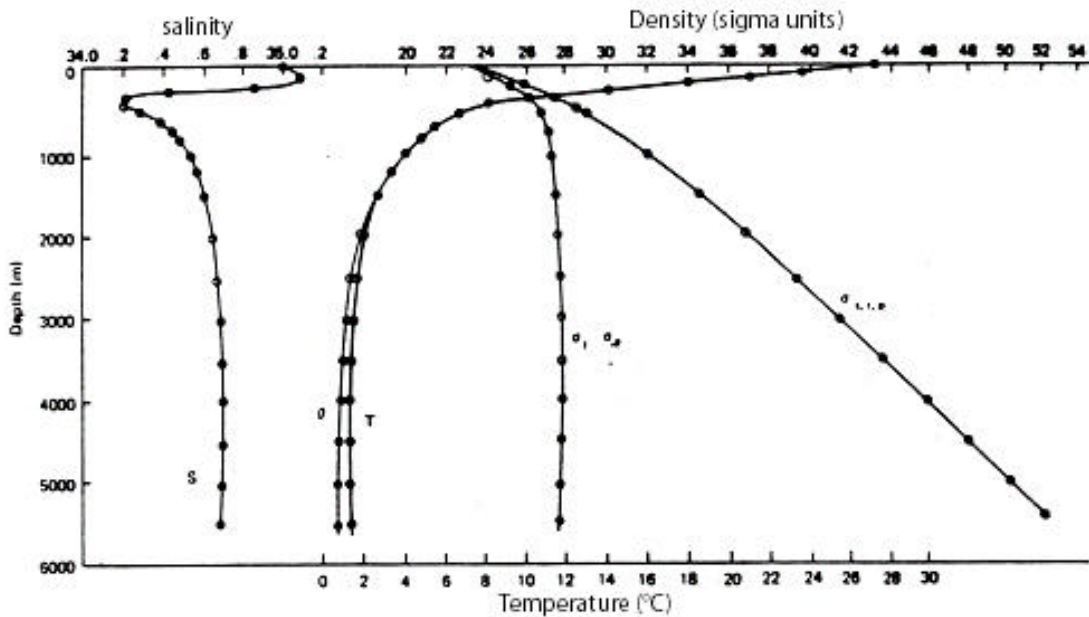
For computing of ocean currents from water properties, it is convenient to use the inverse of density or specific volume according to

$$a_{STP} = \frac{1}{r_{STP}}.$$

A related quantity, known as the specific volume anomaly,  $d$ , is defined

$$d = a_{STP} - a_{35,0^{\circ}C,P},$$

where  $a_{35,0,P}$  is the specific volume of a “standard ocean” that is defined so  $d$  is normally a positive value.  $d$  is somewhat analogous to  $s$ .



**Figure 3.4** Comparison of insitu temperature and potential temperature profiles (left); sigma-t and sigma- $\tau$  profiles.(Knauss, 1978)

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For computational purposes  $d$  can be expressed as follows

$$d = d_s + d_T + d_{ST} + d_{SP} + d_{TP} + d_{STP}$$

$\Delta_{S,T}$   
Thermosteric  
Anomaly  
(Table 11-2)

remaining  
pressure  
effect  
(Table 11-3)

very small  
usually  
ignored

where these components can be obtained from the indicated tables in Knauss. The oceanic range of  $d$  is 50 - 100 x 10<sup>-5</sup> or 50-100 cl/ton, where  $d$  units are

$$[d] = \text{cm}^3/\text{gm} \text{ or } d/\text{ton} = \frac{10 \text{ ml}}{10^3 \text{ kg}}$$

### WATER COLUMN STABILITY

Normally density increases with depth because of the natural tendency for more dense water parcels to sink quickly to their equilibrium depths water parcels which are displaced vertically from this stable configuration will oscillate and eventually return to their equilibrium depth. The strength of the stability is related to the vertical potential density gradient

$$\frac{\partial \mathbf{r}_{\text{pot}}}{\partial z},$$

where

$$\mathbf{r}_{\text{pot}} = 10^{-3} \mathbf{S}_q + 1.$$

A measure of the strength of stability is the natural frequency at which a water parcel would oscillate vertically in the ocean. This is called the buoyancy or Brunt-Väisälä frequency, which varies with depth according to

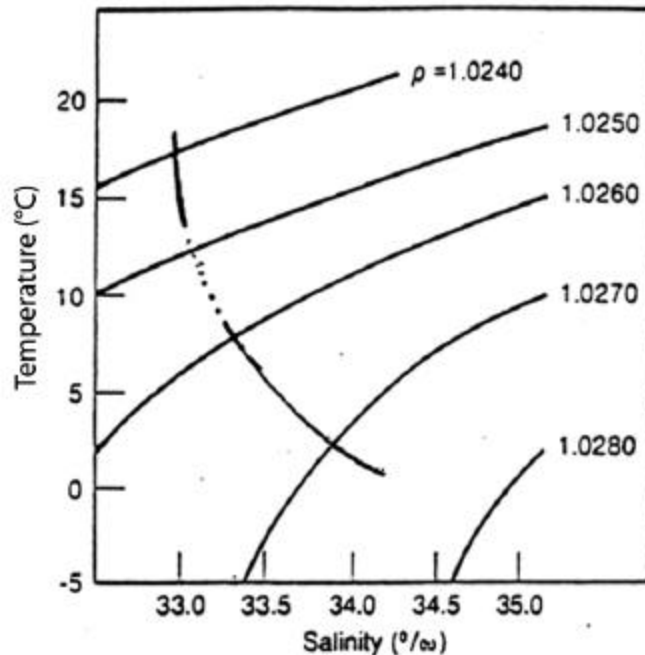
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$$N^2(z) = \left( -\frac{g}{r_o} \frac{\partial r_{pot}}{\partial z} \right); [N] = \frac{\text{rad}}{\text{sec}} \text{ or } E' \cong -\frac{\partial s_q}{r \partial z} \times 10^{-3}$$

Thus, it is the slope of the vertical distribution of potential density anomaly,  $s_q$ , which can be used to assess the stability of the water column or its resistance to vertical motion. For shallow depths, where temperatures and salt effects dominate pressure effect, profiles of  $s_T$  are adequate for assessing water column stability.

### TEMPERATURE AND SALINITY DIAGRAMS

A useful tool in assessing the stability of a particular water column is a T (or  $q$ ) - S diagram; on which T (or  $q$ ) and S measurement pairs are plotted. Because  $s_T$  (or  $s_q$ ) is a unique function of T( $q$ ) and S, isopleths of  $s_T$  (or  $s_q$ ) can be inscribed on the T( $q$ ) - S diagram as (Figure 3.5).



**Figure 3.5.** The density of seawater, abbreviated as  $\rho$ , varies with temperature and salinity. The pressure effect is not included here. (Duxbury & Duxbury, 1984)



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One of the important consequences of the nonlinearity of the  $s_T(s_q)$  isopleths is that

$$\text{near } 28^\circ\text{C } \left. \frac{ds_T}{dT} \right|_{35\%} > \left. \frac{ds_T}{dT} \right|_{35\%} \text{ near } 6^\circ\text{C}.$$

This nonlinear feature of density plays an important role in polar water formation to be discussed later. (The sensitivity of  $s_T$  to salinity changes does not vary much from high to low temperatures.) With our crude understanding of the important air/sea interaction processes and some exposure to the terminology used by oceanographers we can begin to explore the distributions of density, temperature and salinity.

### Surface Distribution of Density

Because the surface ocean temperatures and salinities are most directly influenced by air/sea interaction processes, it is not surprising that the north-south horizontal gradients of density (and temperature) are the largest. Thus the latitude variation of  $s_T$  (Figure 3.6) shows that density ( $s_T$ ) is primarily determined by temperature variation - except in the polar regions where  $ds_T/dT$  is relatively less and salinity variations are significant. Also note that the minimum in  $s_T$  is located at a latitude to about  $5^\circ\text{N}$  (corresponding to the maximum in  $T$  - the “meteorological equator”).

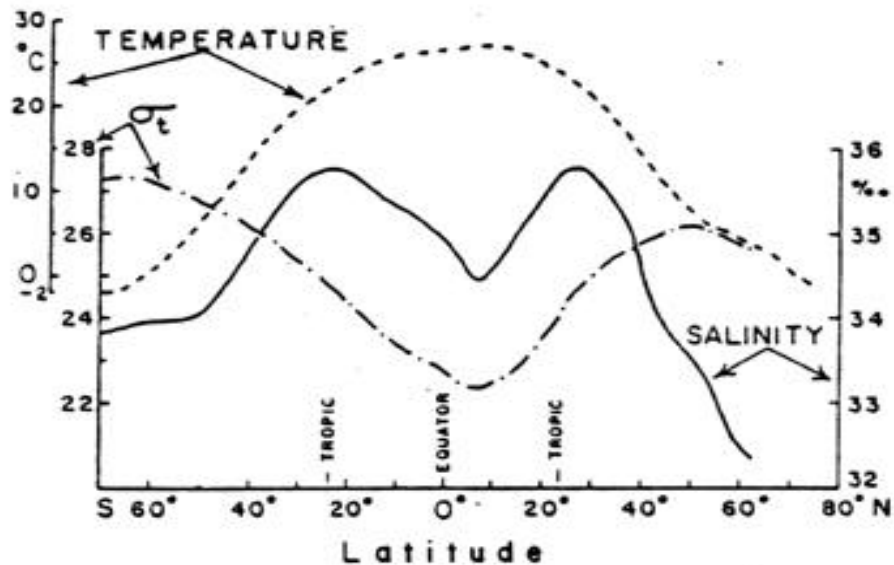


Figure 3.8 Latitudinal variation of average ocean surface temperature, salinity and density ( $s_T$ ).

### Horizontal and Vertical Distributions of Temperature

The zonal distributions of surface temperature (and density) in summer (Figures 3.7) and winter (Figure 3.8) are consistent with the strongly zonal distribution of atmospheric circulation.

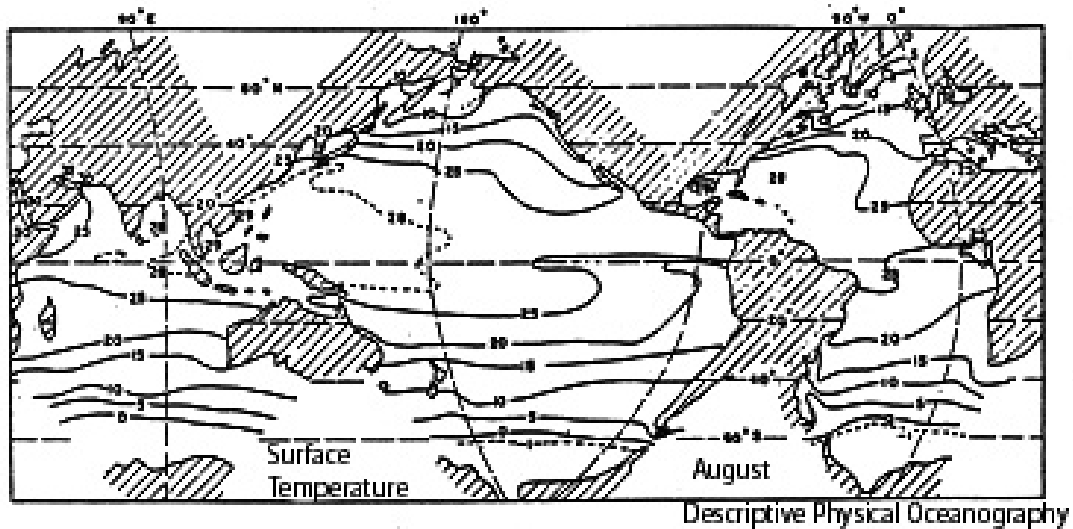


Figure 3.7 Surface temperature of the oceans in August. (Pickard & Emery, 1982)

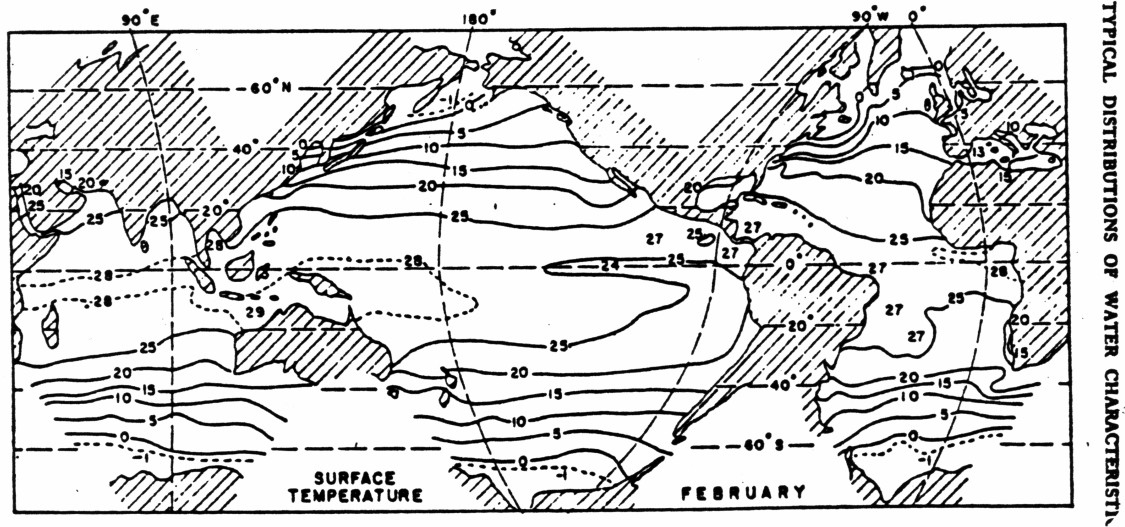


Figure 3.8 Surface temperature of the oceans in February. (Pickard & Emery, 1982)

One of the principal features of the oceans is its vertical stratification of properties. For example vertical gradients of temperature are 5000 times greater than the horizontal

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gradients discussed above. At the equator  $T_{\text{surf}} = 25^{\circ}\text{C}$  and  $T_{1000\text{m}} = 5^{\circ}\text{C}$ , thus vertical temperature gradient at the equator is  $\left. \frac{dT}{dz} \right|_{\text{equ}} = \frac{20^{\circ}\text{C}}{\text{km}}$  (Figure 3.9). On the other hand one must move 5000 km northward on the surface before  $T_{\text{surf}} = 5^{\circ}\text{C}$ . Thus the northward temperature gradient at the ocean surface

$$\text{is } \left. \frac{dT}{dy} \right|_{\text{surf}} = \frac{20^{\circ}\text{C}}{5000 \text{ km}} = 4 \times 10^{-3} \frac{^{\circ}\text{C}}{\text{km}}.$$

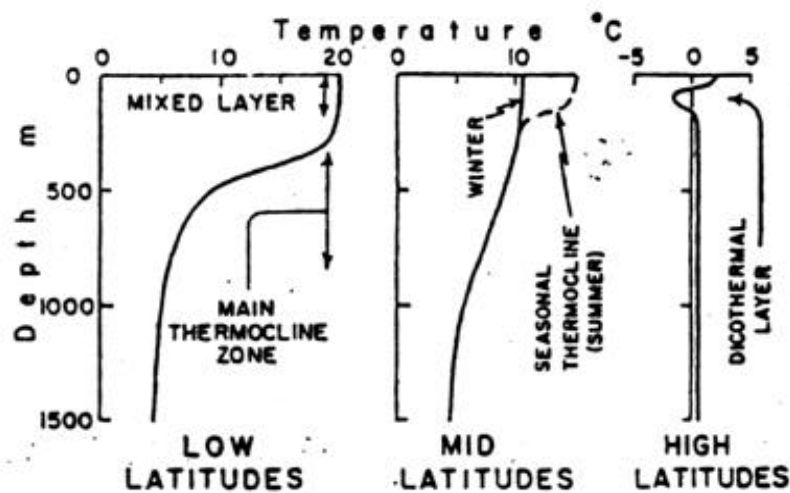
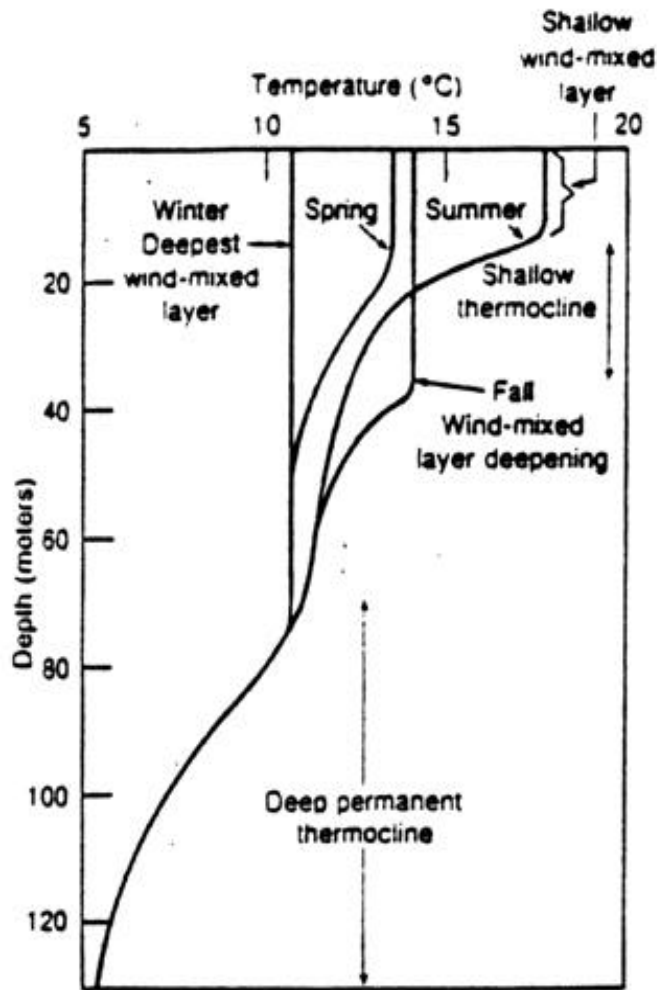


Figure 3.9. Typical mean temperature profiles in the open ocean. (Pickard & Emery, 1982)

### Seasonal Temperature Distributions

The seasonal variability of vertical temperature stratification is summarized in Figure 3.10. The four general zones that define these profiles are described next.

- (1) The mixed layer is found between the surface and 20-60 m depths and strongly reflects the influence of surface winds and heat exchange processes.



**Figure 3.10.** The seasonal variability of a typical subtropical temperature profile. (Duxbury & Duxbury, 1984)

- (2) The seasonal thermocline is usually found in the upper 100 m and is a feature which will change seasonally in intensity. Typical vertical gradients:

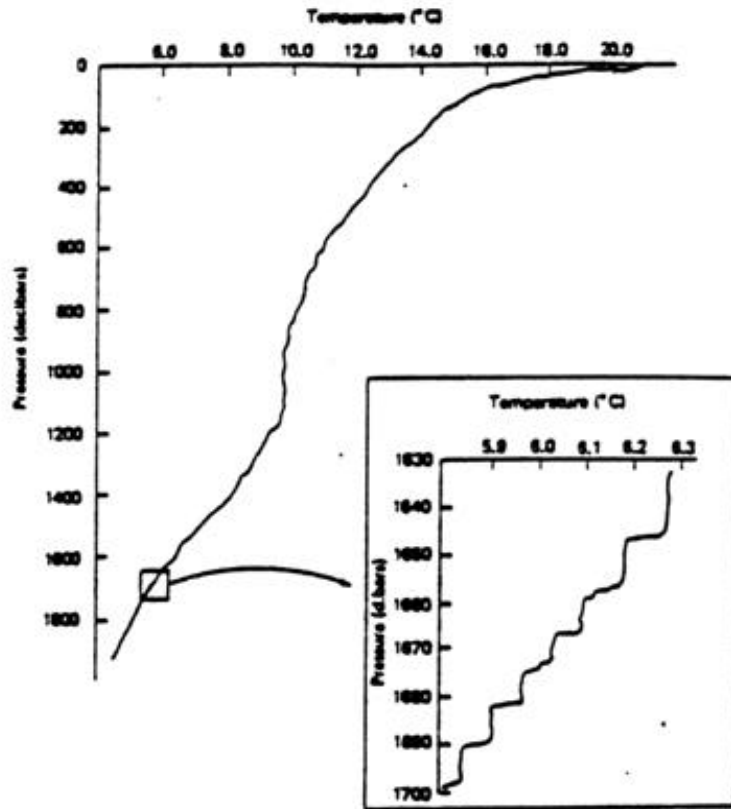
$$\frac{dT}{dz} = 0.06^{\circ}\text{C}/\text{m} \text{ (} 60^{\circ}\text{C}/\text{km} \text{) change with seasons.}$$

- (3) The main or permanent thermocline is found between 800-1500 m and is maintained by the general circulation. Typically  $\frac{dT}{dz} = 0.222 \frac{^{\circ}\text{C}}{\text{m}}$  or  $\frac{22^{\circ}\text{C}}{\text{km}}$  for this region.

- (4) The deep water is nearly isothermal and has typical  $\frac{dT}{dz} = 0.0008 \frac{^{\circ}\text{C}}{\text{m}}$  or  $\frac{0.8^{\circ}\text{C}}{\text{km}}$ .

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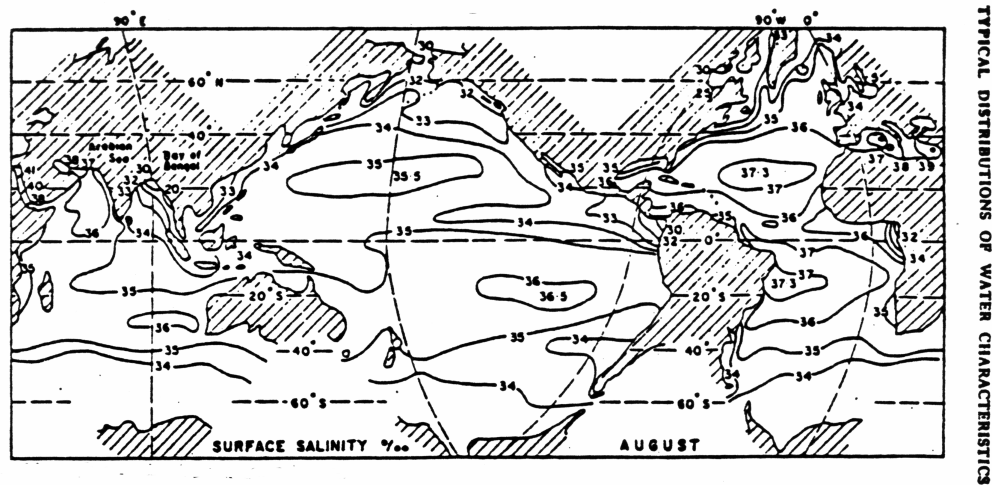
It is important to realize that the “smoothly-varying” profiles of temperature profile, like the one to the left in Figure 3.11, is really highly structured as depicted to the right in Figure 3.11. The rather large “steps” in the temperature trace in are due to small scale mixing processes discussed later.



**Figure 3.11.** Sharp stepwise temperature gradient can be easily seen on the blown-up section of the temperatures traced from the Mediterranean Outflow Region (34°N, 11°W). The temperature steps are generally less than 0.1°C and occur at about 10-m intervals. (After Magnell, B., 1976a “Salt Fingers in the Mediterranean Outflow Region (34°N, 11°W) using a Towed Sensor”, *J. Phy. Oceanogr.*, 6.) (Knauss, 1978)

### Surface Distribution of Salinity

The ocean surface distributions of salt are less zonal and less variable than ocean surface temperature because salt variability is controlled by the spatial variability in the evaporation, precipitation and coastal freshwater outflow. Global scale surface salinity distributions are presented in Figure 3.12.



**Figure 3.12** Surface salinity of the oceans in August.

The lowest salinity 0 to 30 psu are found in estuaries and polar environments near glaciers. Mid-range salinities (30psu-34psu) are typically found in coastal regimes. Typical oceanic salinities range from 33psu to 37psu with Atlantic salinities being slightly higher than Pacific salinities due to the Mediterranean outflow. Semi-enclosed regions, such as the Mediterranean and Red Sea, exhibit salinities of 37psu to 41psu respectively.

### Vertical Profiles of Salinity

Like temperature, oceanic salinities are also vertically stratified because their distribution is strongly influenced by air/sea interaction processes at the sea surface plus the deep circulation in the ocean (see [Figure 3.13](#)). Unlike temperature it is more difficult to generalize about the vertical distribution. Because fresh and salt water sources are not distributed zonally, like the heat sources, the salt distributions are less easily summarized.

In general, however,

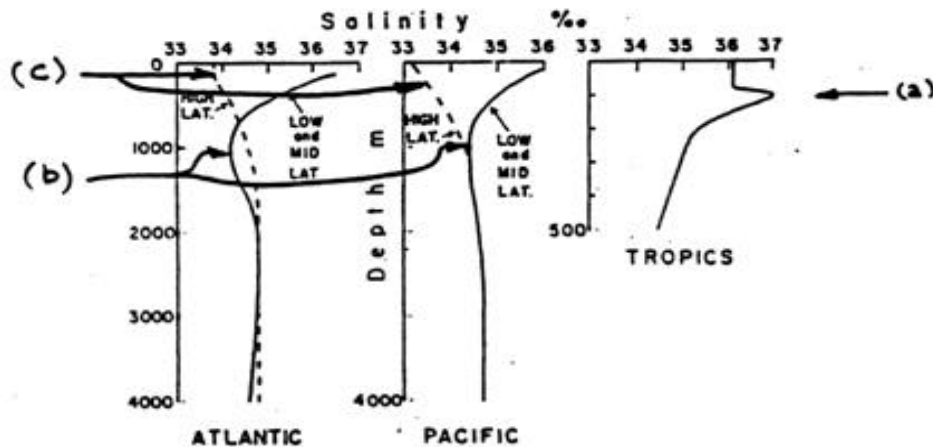
- (a) A salinity maximum occurs at about 100 m depth in the tropics and is due to equatorial circulation patterns (which we will discuss later),

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(b) A salinity minimum is found between 600 m and 1000 m in the equatorial tropics and subtropical latitudes and is due to circulation and mixing,

(c) At high latitudes there is a tendency toward low surface salinity.

Low and mid-latitude profiles, which exhibit strong vertical salinity gradients or haloclines in the upper 800 m, are shown in [Figure 3.13](#). Note that these profiles have a destabilizing influence on the water column. However, temperature effects are predominant in these regions. When salt destabilizes the water column small scale processes which lead to the structure seen in [Figure 3.13](#) can result.



**Figure 3.13** Typical mean salinity profiles in the open ocean. (Pickard & Emery, 1982)

### Vertical Distribution of Density

Because ocean density variation is determined principally by ocean temperature, density exhibits the same kind of stratification as temperature (see [Figure 3.14](#)) - principal stratification in the upper two kilometers of the ocean. However, as shown by the vertical section of **isotherms** in [Figure 3.15](#), the strength of the stratification decreases as one moving poleward from the tropics.

The resistance to vertical motion and mixing increases as the density

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gradient  $-\frac{\partial r_{\text{pot}}}{\partial z}$ , or  $-\frac{\partial s_q}{\partial z}$ , or  $N = \sqrt{-\frac{g}{r} \frac{\partial r_{\text{pot}}}{\partial z}}$  increases (remember z is positive

upwards). The regions of maximum stability within the water column are called pycnoclines. Because stable vertical stratification resists the vertical motion of water, there is a strong tendency for water parcels to move along isopycnals or isopleths of density ( $s_T$  or  $s_q$ ). The vertical section of density in [Figure 3.15](#) suggests a polar origin to the deep water found at mid-latitudes and equatorial regions.

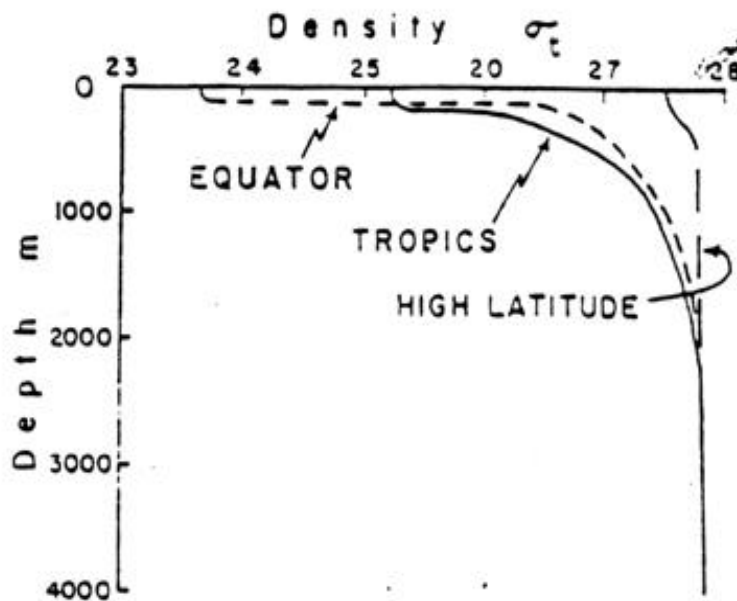
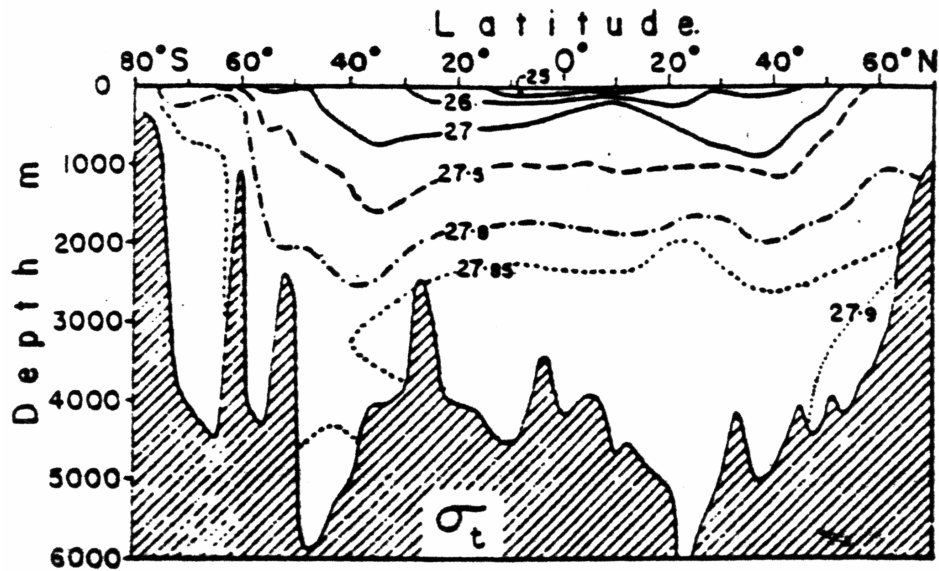


Figure 3.14. Typical density/depth profiles at low and high latitudes. (Pickard and Emery, 1982)





**Figure 3.15** Density ( $\sigma_T$ ) in a south-north section of the western Atlantic (after Wüst). (Pickard & Emery, 1982)

### Sound Transmission in the Sea

The atmosphere transmits light (and other electromagnetic radiation) much better than it transmits sound. While being relatively opaque to electromagnetic radiation transmission, the ocean is transparent to sound transmission. Thus sound has become an important tool for studying the ocean. Difficulties arise however because the profiles of sound speed vary with depth in a number of different ways. As Figure 3.16 shows the sound transmission paths vary also.

Of particular interest here is understanding how the distribution of the speed of sound,  $C_s$  varies with ocean properties. Like other physical properties of the ocean the speed of sound depends upon the local values of temperature, salinity and pressure. It can be expressed symbolically as follows

$$C_s = C_s(T, S, P),$$

where  $T = q + T_A$

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with  $T_A$  being the part of  $T$  due to adiabatic compression.

(i.e.  $dT_A/dz \sim -0.1 \text{ }^\circ\text{C/km}$ )

Therefore we can write

$$C_s = C_s(\mathbf{q}, S, T_A)$$

By taking the differential of this relation and forming a quantity we define as the fractional change of speed of sound with depth or

$$\frac{1}{C_s} \frac{\partial C_s}{\partial z} = \mathbf{a} \frac{\partial \mathbf{q}}{\partial z} + \mathbf{b} \frac{\partial S}{\partial z} + \mathbf{g}_A$$

where  $\mathbf{g}_A$  is the fractional change of the speed of sound in an adiabatic ocean and  $\mathbf{a}$ ,  $\mathbf{b}$  are constants.

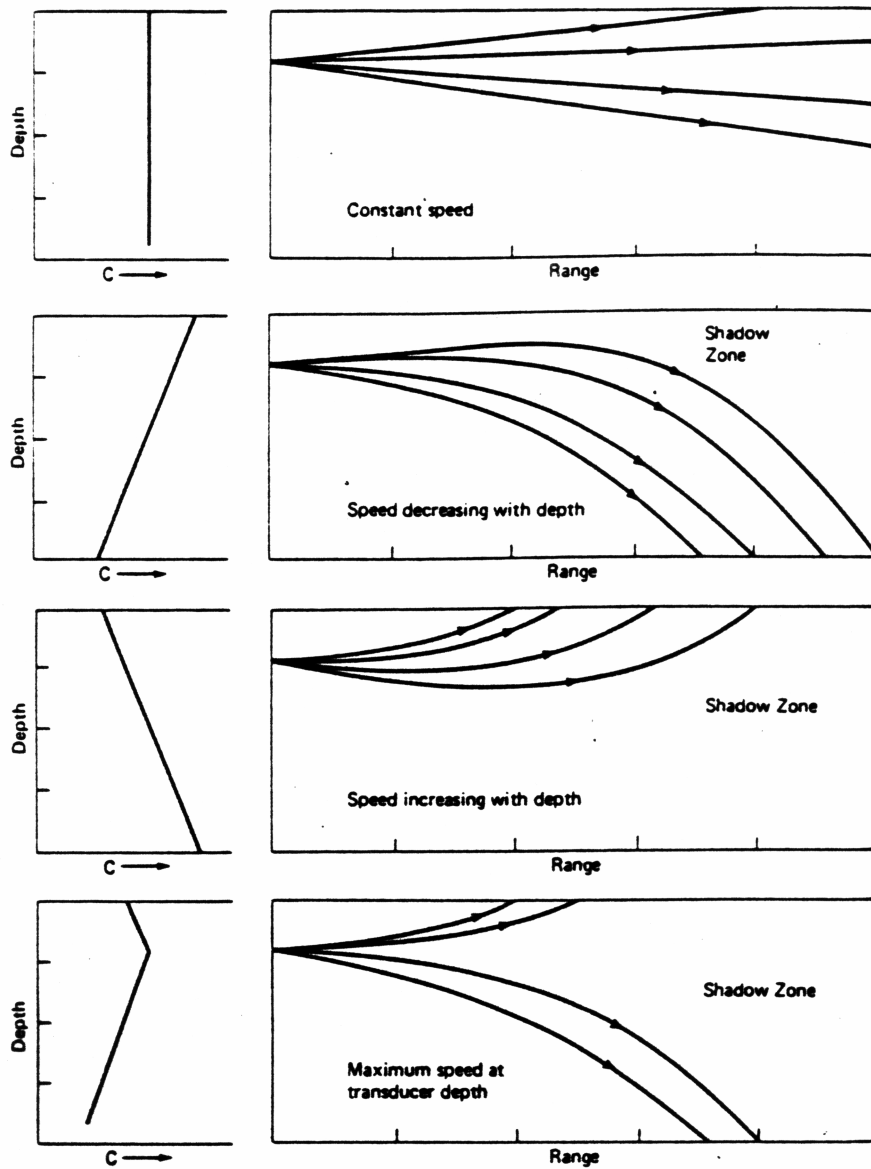
We also know that

$$\mathbf{r} = \mathbf{r}(T, S, P) \text{ or } \mathbf{r}_{\text{pot}} = \mathbf{r}_{\text{pot}}(\mathbf{q}, S)$$

Therefore the fractional change of  $\mathbf{r}_{\text{pot}}$  with depth is

$$\frac{1}{\mathbf{r}_{\text{pot}}} \frac{\partial \mathbf{r}_{\text{pot}}}{\partial z} = \mathbf{a} \frac{\partial \mathbf{q}}{\partial z} + \mathbf{b} \frac{\partial S}{\partial z},$$

where  $\mathbf{a}$  and  $\mathbf{b}$  are constants.



**Figure 3.16.** If the sound velocity is constant, the sound travels in straight lines; if the sound velocity changes, the sound rays are refracted. The direction is always toward lower velocities. (Knauss, 1978)

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The differential relations for  $C_s$  and  $r_{\text{pot}}$  depend on the vertical gradients of  $q$  and  $S$  and can therefore be combined. Without the gory detail the result is a relation between fractional sound speed change and the square of the Brunt-Viäsälä frequency plus the compressional effects of pressure,

$$\frac{1}{C_s} \frac{\partial C_s}{\partial z} = -\left(\frac{\rho}{\rho_0}\right) N^2(z) + g_A$$

(a)

Term a contains the effects of the vertical density gradient which is due to changes in the intrinsic properties of  $q$  and  $S$ .

(b)

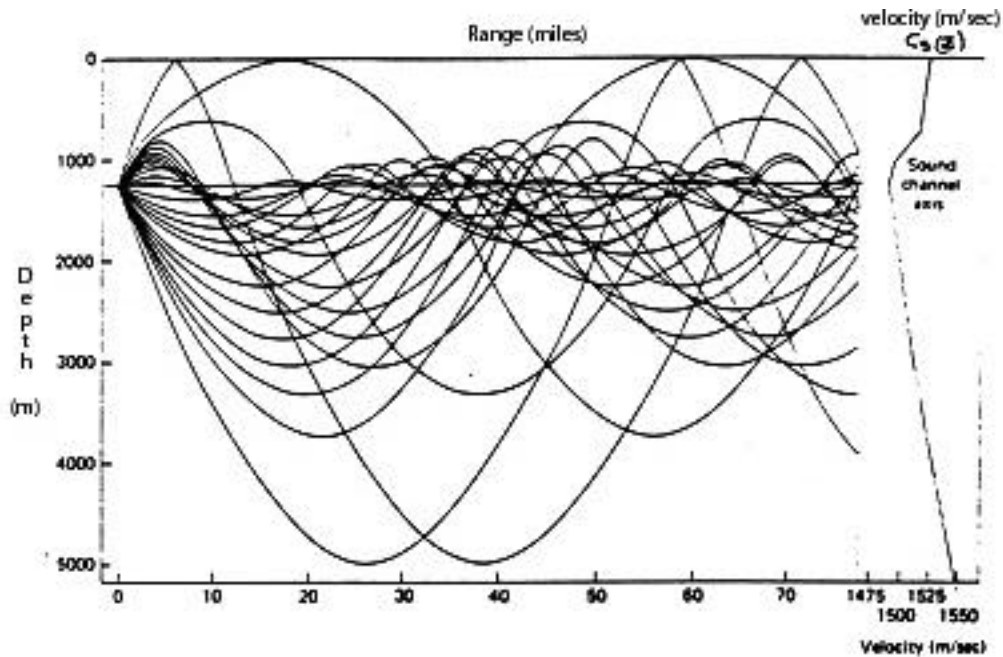
Term b contains the effects on sound speed due to the external effects of pressure.

We can solve for  $C_s(z)$  from the relation above (Figure 3.17 and Appendix B).

There are several observations we can make about the shape of this curve; namely

- (a) The speed of sound decrease with depth above 1.2 km due to dominant effects of  $q$  and  $S$ .
- (b) The speed of sound increases with depth below 1.2 km due to the dominance of pressure effects (in the absence of  $q$ ,  $S$  variability).

The minimum sound speed occurs at a depth near the depth of the main pycnocline. The consequences of this sound speed distribution are important for the way sound travels in the ocean. We can explore the important features by considering what happens when we explode a depth charge at the depth of the minimum sound speed (~1.2 km depth)  $C_{s1}$ . The acoustic energy radiates spherically (locally) in the region of the charge.



**Figure 3.17.** Because sound rays are always bent toward the lower velocity, a sound minimum tends to trap sound waves and to “channel” them. A sound source in the deep sound channel can often be heard hundreds and even thousands of miles away. The sounds that travel those long distances are of low frequency (the order of 100 to 1,000 c/sec) because low-frequency is less attenuated than high frequency. Although the ocean is reasonably “transparent” to sound waves, it is not completely so, and over a wide range of frequency, attenuation increases as the square of the frequency; thus a doubling of the frequency increases the attenuation by four times, which means that long-range sound transmission is best at low frequencies. Frequencies as low as 100 c/sec (with a corresponding wavelength of about 15 m) are commonly used for very long-range sound transmissions. Fresh water is even more transparent than salt water to the transmission of sound; for a given frequency the attenuation is 100 times less in fresh water. The difference is caused by a chemical interaction of the sound waves with one component of the sea salt, the magnesium sulphate ion. (Knauss, 1978)

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All energy is refracted along “ray paths” governed by Snell’s Law i.e.

$$\frac{C_s}{\cos q'} = \frac{C_{s1}}{\cos q_1'} = \text{constant} ,$$

where  $q'$  is the elevation above horizontal for the particular ray of interest.

(Choose as initial angle  $q_1'$ , then Snell’s Law governs the angle of the ray,  $q'$ , as the energy propagates into regions of  $C_s(z)$ ).

The only energy to be transmitted horizontally is contained along ray paths between the surface limited ray (SLR), the bottom limited ray (BLR) and the horizontal. Energy travels fastest along the SLR/BLR ray and slowest along the horizontal. However all energy in this “envelope” eventually focuses on the convergence zone at the depth of the minimum sound speed about 50 km “downstream”. Since there is a tendency for sound concentration (as different rays crisscross) along the depth of minimum sound speed the region is referred to as the “sound channel”.

It turns out that in practice the effective depth range of the sound focusing properties of the ocean is from 1 to 2 km and its existence is used to track SOFAR floats, which are ballasted to be neutrally buoyant and transmit only a few watts of acoustic energy over distances of 10,000 km.

A more recent application - acoustic tomography uses differences in sound transmission times to infer changes in ocean property and current structure. One such experiment seeks to monitor the long term change in the average temperature of the ocean. [Figure 3.18](#) summarizes a pilot study for such an experiment.

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What's the Sound of One Ocean Warming?

Oceanography has made a noises in the Indian Ocean ("heard" in Bermuda!) that could be used to measure global warming.

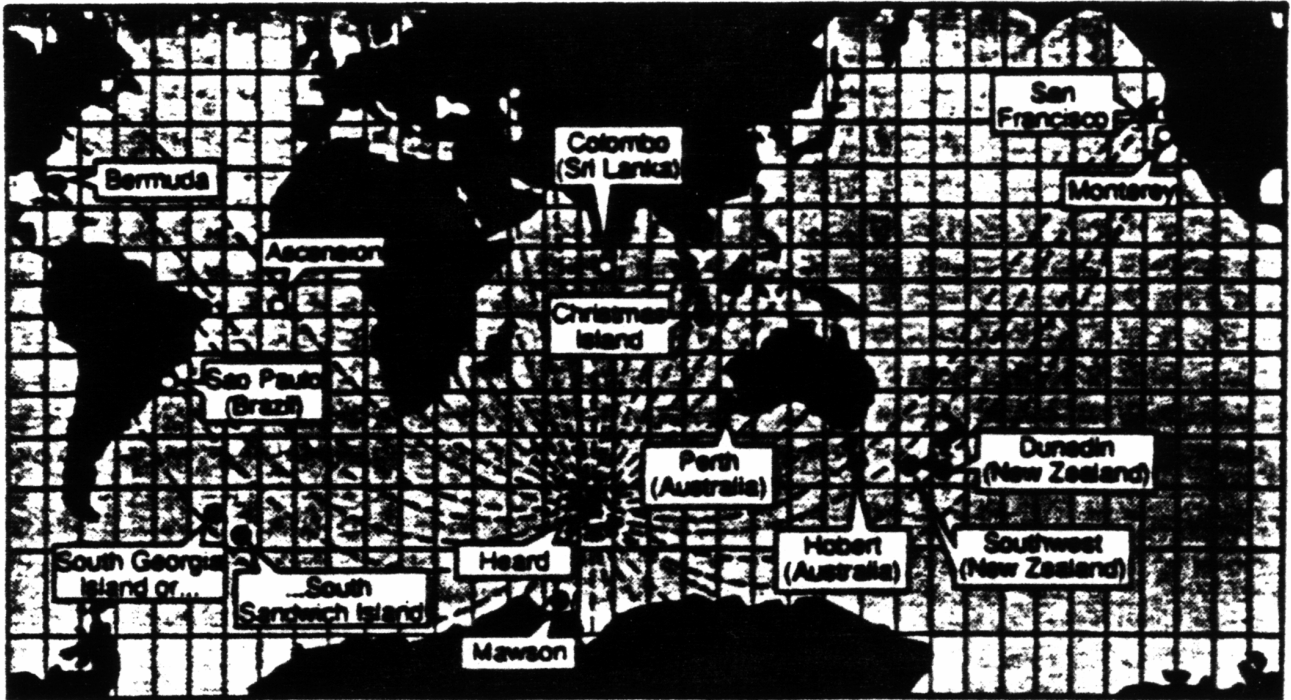


Figure 3.18. Straight shot. Sound waves generated at Heard Island may be detected by hydrophones at Bermuda and San Francisco, as well as other

Chapter 3 - Appendix: Sea Water Density Terminology				
Quantity	Units	Accuracy of Measurement		Typical Range
		<u>Classical</u>	<u>Modern</u>	
T Temperature	$^{\circ}\text{C} = \frac{5}{9} (^{\circ}\text{F} - 32)$	0.02 $^{\circ}\text{C}$	abs: 0.005 $^{\circ}\text{C}$ diff: 0.001 $^{\circ}\text{C}$	-0.2 to 30 $^{\circ}\text{C}$
S Salinity	gm/kg or ‰	0.02 ‰	0.0024 ‰	33-37 ‰
Cl Chlorinity	gm/kg			
P Pressure	decibars = $10^5$ dynes/cm <sup>2</sup> [weight of 1m of sea water]	abs. 25 db	abs. 1.5 db diff. 0.01 db	0.6000 db
$\rho_{STP}$ Density	gm/cm <sup>3</sup>	$10^{-5} \frac{\text{gm}}{\text{cm}^3}$	$10^{-6} \frac{\text{gm}}{\text{cm}^3}$	1.03 to 1.07 $\frac{\text{gm}}{\text{cm}^3}$
$\sigma_{STP}$ Density anomaly $= (\rho_{STP}^{-1}) \times 10^3$		$10^{-2}$	$10^{-3}$	30-70
$\rho_{ST0}$ Density at atmospheric pressure	gm/cm <sup>3</sup>	$10^{-5} \frac{\text{gm}}{\text{cm}^3}$	$10^{-6} \frac{\text{gm}}{\text{cm}^3}$	1.023-1.028 $\frac{\text{gm}}{\text{cm}^3}$
$\sigma_T$ Density anomaly at atmospheric pressure-“Sigma-T” $= (\rho_{S,T,0}^{-1}) \times 10^3$		$10^{-2}$	$10^{-3}$	23-28
$q$ Potential temperature	? C	0.02? C	abs: 0.005 $^{\circ}\text{C}$ diff: 0.001 $^{\circ}\text{C}$	-0.2-30 $^{\circ}\text{C}$
$\rho_{sqP} = 0$	gm/cm <sup>3</sup>	$10^{-2}$	$10^{-3}$	23-28
$\sigma_q$ Potential density anomaly	$10^{-3} \frac{\text{gm}}{\text{cm}^3}$			
$d_{STP}$ Specific volume	cm <sup>3</sup> /gm (or ml/gm)	$10^{-5}$	$10^{-6}$	0.02-0.97
$d = d_s + d_T + d_{ST} + d_{SP} + d_{STP}$ Specific volume anomaly $(a_{STP} - a_{35,0P})$	$\frac{\text{cl}}{\text{ton}} \left( = \frac{10 \text{ ml}}{10^3 \text{ kg}} \right)$	$10^{-5}$	$10^{-6}$	0.02-0.97
$\Delta_{ST} = d_s - d_{TP} - d_{SP}$ Thermosteric anomaly				



Chapter 3 - Appendix: Other Physical Properties of Sea Water

Quantity	Definition	Units	Typical Values
K Compressibility	$\frac{-1}{\alpha_{STP}} \frac{\partial \alpha_{STP}}{\partial p}$	$\frac{\text{sec}^2 - \text{cm}}{\text{gm cm}}$	$4 \times 10^{-5}$
$\beta$ Coefficient of Thermal Expansion	$\frac{1}{\alpha_{STP}} \frac{\partial \alpha_{STP}}{\partial T}$	$\frac{1}{^\circ\text{C}}$	$10^{-4}$
$\mu$ Molecular Viscosity	constant of proportionality between shear stress and velocity gradient	$\frac{\text{gm}}{\text{cm- sec}}$	0.02
$\nu$ Kinematic Viscosity	$\mu / \rho$	$\frac{\text{cm}^2}{\text{sec}}$	0.02
k Thermal Conductivity	constant of proportionality between heat flux and temperature gradient	$\frac{\text{cal}}{\text{cm- }^\circ\text{C- sec}}$	$10^{-3}$
Cp Specific Heat (at constant pressure)	amount of heat per gm to warm substance 1 °C	$\frac{\text{cal}}{\text{gm }^\circ\text{C}}$	1
L Latent Heat of Evaporation	amount of heat to evaporate one gm of substance	$\frac{\text{gm- cal}}{\text{gm}}$	600
Surface tension	$75.64 - 0.144 + 0.399Cl$	$\frac{\text{dyne}}{\text{cm}^2}$	75
Vs Speed of Sound	$= \frac{1}{\rho_{STP} K}$	$\frac{\text{cm}}{\text{sec}}$	$1.5 \times 10^5$
C Speed of Light	---	$\frac{\text{cm}}{\text{sec}}$	$2.3 \times 10^{10}$
n Refractive Index	---	---	1.3

## Chapter 3 - PROBLEMS

## Problem 3.1. Water Properties

a) Sea Water Density

The following table lists the measured values of pressure  $P_{\text{insitu}}$  (decibars or dbars), temperature  $T_{\text{insitu}}$  ( $^{\circ}\text{C}$ ) and salinity  $S$  ( practical salinity units or psu) at an oceanic hydrographic station.

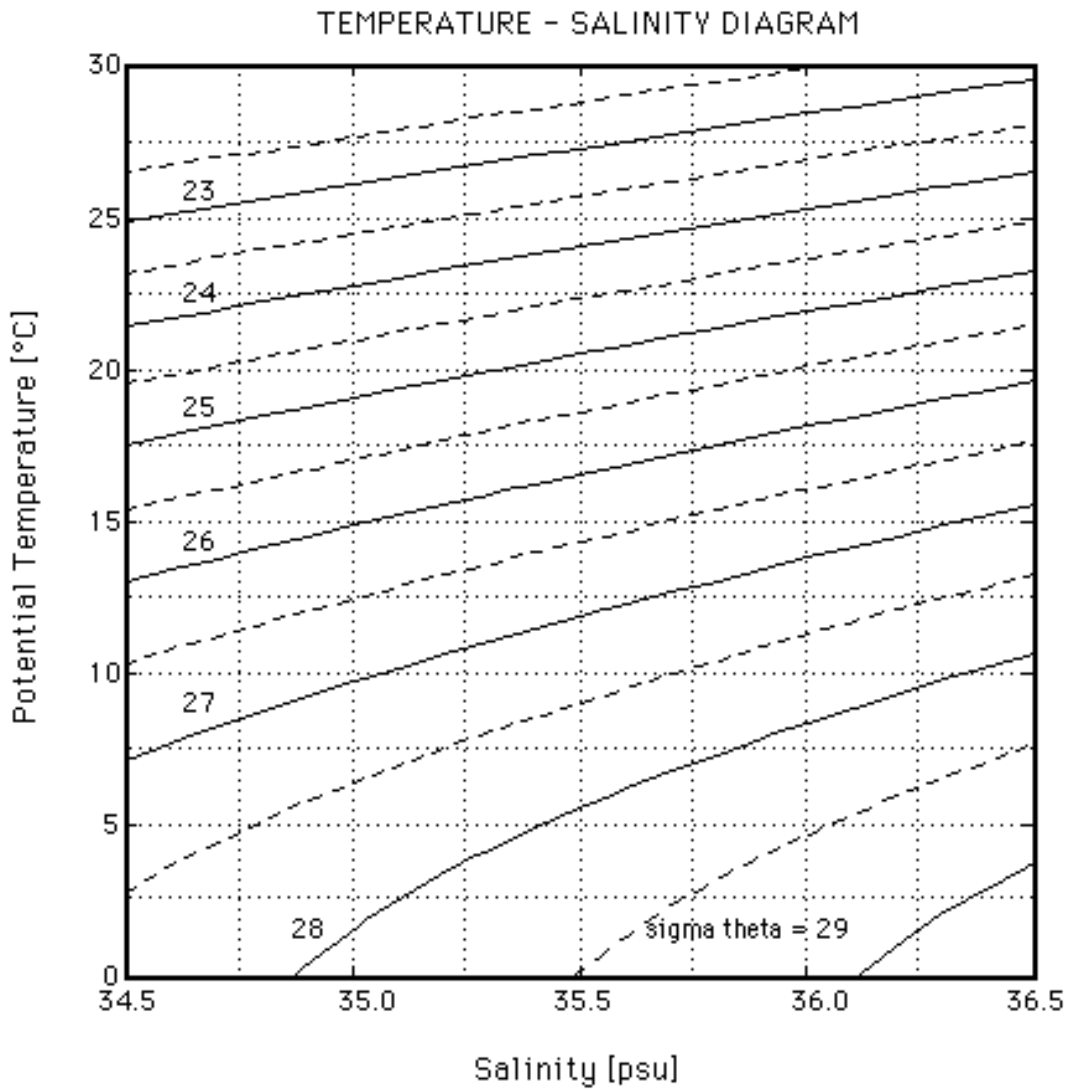
Produce the  $T_{\text{insitu}}$ - $S$  curve for these data on the form provided below.

$P_{\text{insitu}}$ (dbars)	$T_{\text{insitu}}$ ( $^{\circ}\text{C}$ )	$q$ ( $^{\circ}\text{C}$ )	$S$ (psu)	$s_T$	$s_q$	$\Delta_{TS}$	$\delta_{PT}$	$\delta_{PS}$	$\delta$	$\alpha_{35,o,p}$	$\alpha$	$\rho$
0	24.84		35.031									
180	13.01		35.291									
740	5.01		34.488									
1505	4.15		34.950									
5030	1.10		34.754									
7020	1.37		34.755									

b) Potential Temperature

Assuming that the adiabatic temperature gradient  $dT/dP$  in the ocean is a constant  $0.14^{\circ}\text{C}$  per 1000 decibars, estimate the potential temperature  $q$  of the water at the different temperature  $T_{\text{insitu}}$  and pressure  $P_{\text{insitu}}$ . (Hint: Suppose  $T(P) = AP + B$ . What are A and B? Remember that  $T(P = 0) = q$ ).

c) Now plot the  $q$  - $S$  curve on the form below.



- d) Estimate the rest of the information in the table, including the potential density anomaly  $\sigma_q = [\rho(S, T = q, P = 0) - 1] \times 1000$ , from the oceanographic tables provided. Are the density anomaly values consistent with those on the T-S and  $q$ -S diagrams?
- e) Using your new knowledge of global ocean water properties “guess” the region of the world’s oceans where this hydrocast was made?

**Problem 3.2. Oceanographic Vertical Sections**

- a) On page 33, you are provided with an array of measured temperature distributions in an unknown coastal zone. On page 32 you are provided with a sub-sampled (i.e. under-sampled) version of the page 33 temperatures. Contour the page 32 and page 33 vertical sections of temperature data using 6.0, 6.5, 7.0, 7.5, 8.0, and 8.5°C isotherms where appropriate. Make a guess as to what processes may be affecting the observed temperature distribution.
- b) On page 34, find the measured salinity distribution based on several Nansen bottle casts. Contour these results in terms of 33.20, 33.40, 33.60, 33.80 psu isohalines.
- c) On the page 35, compute all possible  $\sigma_T$  values and then contour the isopycnals starting with  $\sigma_T = 26.0$  at  $\Delta\sigma_T = 0.1$  increments.
- d) Now plot T-S diagrams for the 7 profiles data. What can you say about the static stability of each of the profiles? (Hint: Sketch isopycnals on T-S diagram.)

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Temperature (° C) Section - I

0	6.4	6.6	7.3	7.4	7.5	7.8	7.9	7.8	7.6	7.9	7.8	8.1	8.2	
20	°	°	°	°	°	°	°	°	°	°	°	°	°	
40	•7.0	°	•7.7	°	•7.8	°	•8.1	°	•7.9	°	•7.9	°	•8.2	
60	°	°	°	°	°	°	°	°	°	°	°	°	•	
80	•7.8	°	•7.6	°	•7.7	°	•8.2	°	•7.9	°	•7.8	°	•8.3	
100	°	°	°	°	°	°	°	°	°	°	°	°	•	
120			•7.6	°	•7.6	°	•8.0	°	•7.9	°	•8.0	°	•8.4	
140					°	°	°	°	°	°	°	°	°	
160						°	•7.6	°	•8.0	°	•8.5	°	•8.6	
180							°7.6	°7.6			°8.4	°8.6	°8.8	°8.8
200														•9.0

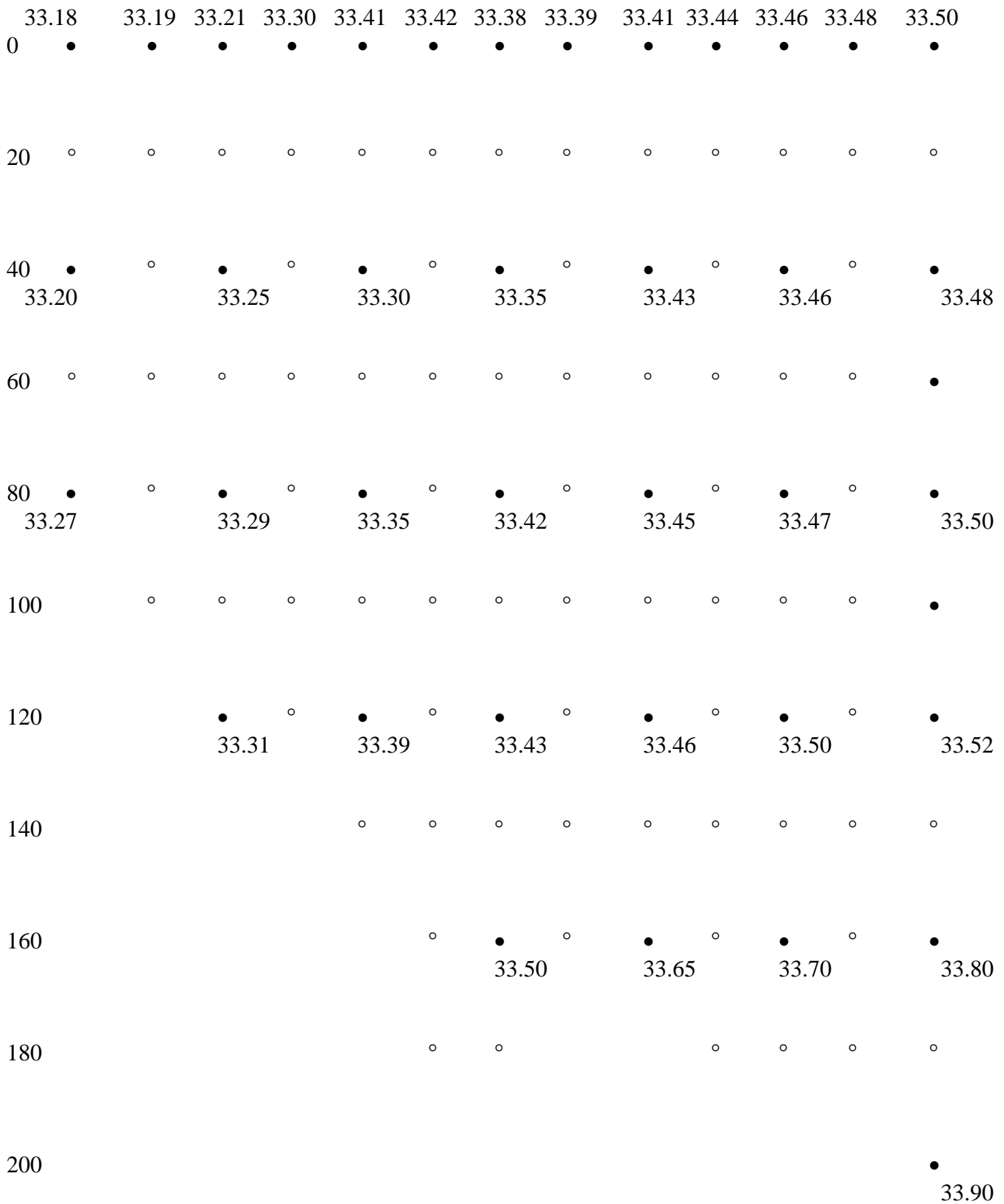
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Temperature (°C) Section - II

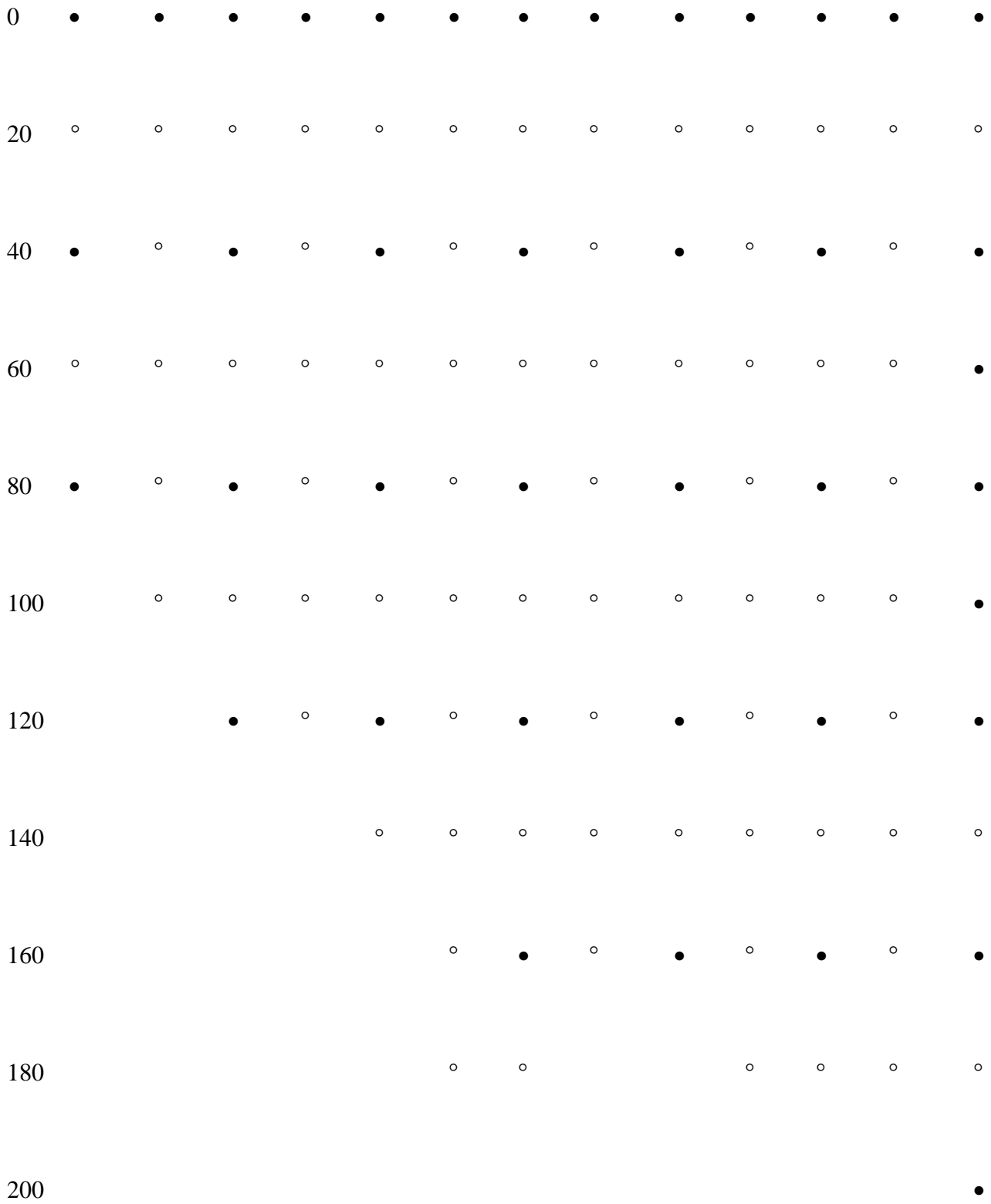
0	6.4	6.6	7.3	7.4	7.5	7.8	7.9	7.8	7.6	7.9	7.8	8.1	8.2
20	°6.5	°7.1	°7.6	°7.7	°7.8	°7.9	°8.0	°7.9	°7.9	°8.1	°7.9	°8.1	°8.2
40	•7.0	°7.5	•7.7	°7.8	•7.8	°7.9	•8.1	°7.9	•7.9	°8.1	•7.9	°8.1	•8.2
60	°7.5	°7.6	°7.6	°7.7	°7.8	°7.9	°8.0	°7.9	°7.9	°8.1	°7.9	°8.1	•8.2
80	•7.8	°7.7	•7.6	°7.7	•7.7	°7.9	•8.2	°7.9	•7.9	°8.2	•7.8	°8.1	•8.3
100		°7.7	°7.6	°7.7	°7.7	°7.9	°8.2	°8.0	°7.9	°7.9	°8.0	°8.1	•8.4
120			•7.6	°7.6	•7.6	°7.9	•8.0	°7.9	•7.9	°8.1	•8.0	°8.2	•8.4
140				°7.6	°7.7	°7.8	°7.7	°7.8	°8.1	°8.2	°8.4	°8.5	
160					°7.6	•7.6	°7.8	•8.0	°8.1	•8.5	°8.6	•8.6	
180					°7.6	°7.6			°8.4	°8.6	°8.8	°8.8	
200													•9.0

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Salinity (psu) Section



Density ( $S_T$  Units) Section





Chapter 3 - pg. 33

Salinity (psu)

33.0

33.2

33.4

33.6

33.8

34.0

10.0

9.0

8.0

C)

7.0

6.0

5.0

**Problem 3.3 Water Column Stability**

Given what you know about properties that affect the density of seawater:

- a) Would coastal regions have a high or low Brunt-Vaisala frequency relative to the central oceanic regions?
- b) Use the simplified definition of static stability to explain your answer.