

JODY M. KLYMAK

INTRODUCTION TO PHYSICAL OCEANOGRAPHY

UNIVERSITY OF VICTORIA

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Contents

	<i>I Overturning circulations</i>	10
1	<i>Estuaries</i>	11
1.1	<i>Pressure differences</i>	12
1.2	<i>Estuarine Flow</i>	13
1.2.1	<i>Salt-wedge estuary</i>	13
1.2.2	<i>Fjord-type estuary</i>	13
1.2.3	<i>Flood-plain estuaries</i>	14
1.2.4	<i>Time-dependence</i>	16
1.2.5	<i>Topographic blocking (mostly in fjords)</i>	19
1.3	<i>Quantifying the Circulation: Knudsen Relation</i>	19
1.3.1	<i>The Conservation of Volume</i>	20
1.3.2	<i>Conservation of salt</i>	20
1.3.3	<i>Application to Estuaries: The Knudsen Relation</i>	21
1.4	<i>Exercise</i>	23
1.5	<i>Example problems</i>	26
2	<i>Equation of State and Pressure Gradients</i>	28
2.1	<i>Temperature</i>	28
2.1.1	<i>Measurement</i>	29
2.1.2	<i>In-situ and potential temperature</i>	30
2.2	<i>Salinity</i>	31
2.2.1	<i>Measurement</i>	32

2.3	<i>Pressure</i>	32
2.3.1	<i>Measuring pressure</i>	33
2.4	<i>Calculating density</i>	34
2.4.1	<i>Density relation</i>	35
2.4.2	<i>Potential density</i>	35
2.4.3	<i>Cabbeling</i>	37
2.5	<i>Other seawater properties</i>	38
2.5.1	<i>Sound</i>	39
2.5.2	<i>Light transmission</i>	39
2.5.3	<i>Ice</i>	40
2.6	<i>Pressure Gradients</i>	40
2.6.1	<i>Hydrostatic Pressure</i>	41
2.6.2	<i>Single density fluid</i>	41
2.6.3	<i>Two-density fluid</i>	42
2.7	<i>Pressure gradients</i>	43
2.7.1	<i>Surface pressure gradients</i>	43
2.7.2	<i>Internal pressure gradients</i>	44
2.8	<i>Exercise</i>	47
2.9	<i>Example problems</i>	47
3	<i>Heat and Salt fluxes</i>	49
3.1	<i>Surface heat fluxes</i>	49
3.1.1	<i>Incoming shortwave radiation</i>	50
3.1.2	<i>Longwave radiation (back radiation)</i>	58
3.1.3	<i>Sensible heat transfer</i>	58
3.1.4	<i>Latent heat losses</i>	59
3.1.5	<i>Integrated Surface Heat Fluxes</i>	61
3.2	<i>Surface freshwater fluxes</i>	62
3.2.1	<i>Evaporation and precipitation patterns</i>	64
3.3	<i>Conservation of mass, contaminants (and heat) in the ocean</i>	66
3.3.1	<i>Conservation of water mass and volume</i>	66
3.3.2	<i>Advective transport and flux</i>	68

3.4	<i>Diffusive fluxes and transports</i>	71
3.5	<i>Net flux: advection plus diffusion</i>	72
3.5.1	<i>Turbulent diffusivities</i>	74
3.5.2	<i>Turbulent diffusion</i>	75
3.6	<i>Exercises</i>	77
3.7	<i>Example problems</i>	79
4	<i>Ocean Stratification and the Mixed Layer</i>	82
4.1	<i>Vertical structure of the ocean</i>	82
4.2	<i>Buoyancy frequency</i>	83
4.3	<i>Overturns</i>	84
4.4	<i>The surface mixed layer</i>	85
4.4.1	<i>Seasonal and diurnal cycles</i>	88
4.4.2	<i>Global mixed layer depths</i>	90
4.5	<i>Restratification of the mixed layer</i>	91
4.6	<i>What maintains the deep stratification?</i>	91
4.7	<i>Exercises</i>	91
5	<i>Ocean water masses</i>	94
5.1	<i>Surface distributions of temperature and salinity</i>	95
5.2	<i>Meridional sections</i>	96
5.3	<i>Vertical structure and θ-S diagrams</i>	99
5.4	<i>Summary</i>	105
5.5	<i>Exercises</i>	107
5.6	<i>Example problems</i>	108
6	<i>Thermohaline overturning circulation</i>	110
6.1	<i>Inferring the overturning circulation from other tracers</i>	111
6.1.1	<i>Bio-active tracers</i>	111
6.1.2	<i>Radioactive and tracers from the crust</i>	114
6.1.3	<i>Man-made tracers</i>	117

6.2	<i>Deep water formation processes</i>	121
6.2.1	<i>Antarctic Bottom Water</i>	121
6.2.2	<i>North Atlantic Deep Water</i>	122
6.3	<i>Inferring the strength of the overturning circulation from data</i>	123
6.3.1	<i>How to make an inverse calculation</i>	127
6.3.2	<i>Inverse results</i>	130
6.4	<i>Theory of overturning circulation and Sandstrom's theorem</i>	133
6.5	<i>Exercises</i>	136
6.6	<i>Example problems</i>	137
7	<i>Climate Change; The Effect on the Overturning Circulation and Sea Level</i>	140
7.1	<i>Effect of melting on the overturning circulation</i>	140
7.2	<i>Sea level change</i>	144
7.2.1	<i>Land-fast ice</i>	145
7.2.2	<i>Thermal expansion (steric change)</i>	146
7.3	<i>Future sea level rise</i>	146
7.4	<i>Other reading</i>	147
II	<i>Waves and Tides</i>	150
8	<i>Surface gravity waves</i>	151
8.1	<i>Characterizing waves</i>	152
8.2	<i>Causes</i>	153
8.3	<i>Dispersion relation, phase speed, and particle motion</i>	155
8.4	<i>Orbital motions: velocity of water parcels</i>	156
8.5	<i>Group speed</i>	158
8.6	<i>Energy conservation</i>	163
8.7	<i>Refraction and diffraction</i>	164
8.8	<i>Non-linearity</i>	166
8.9	<i>Exercises</i>	167

9	<i>Tides</i>	170
9.1	<i>Tidal example and describing tides</i>	170
9.2	<i>Tidal forces: Gravity and centrifugal forces on co-orbiting bodies</i>	171
9.3	<i>The reason for various tidal frequencies</i>	175
9.3.1	<i>Twice a day tides</i>	176
9.3.2	<i>Daily tides (diurnal): Tides due to non-equatorial orbits</i>	178
9.3.3	<i>Longer period tides</i>	180
9.4	<i>Dynamic theory of the tides</i>	181
9.4.1	<i>Tidal Resonance</i>	181
9.4.2	<i>Frictional effects</i>	183
9.4.3	<i>Global response</i>	183
9.5	<i>Exercises</i>	185
9.6	<i>Example problems</i>	186
	 <i>III Wind-driven circulations</i>	 187
10	<i>Wind-driven circulation overview</i>	188
10.1	<i>Coastal upwelling</i>	188
10.2	<i>North Pacific observations</i>	193
11	<i>Coriolis force and Ekman Balance</i>	206
11.1	<i>Quick review of forces and apparent forces</i>	206
11.2	<i>Centrifugal and Coriolis forces</i>	208
11.2.1	<i>Centrifugal force</i>	208
11.2.2	<i>Coriolis force</i>	210
11.3	<i>Friction force</i>	214
11.3.1	<i>External stresses</i>	215
11.4	<i>Ekman balance</i>	216
11.4.1	<i>The Ekman spiral: fact and fiction</i>	218
11.5	<i>Ekman convergence and divergence</i>	220
11.6	<i>Exercises</i>	223
11.7	<i>Example problems</i>	224

12	<i>Geostrophic balance and coastal flow</i>	225
12.1	<i>Geostrophic balance</i>	225
12.1.1	<i>Thermal wind</i>	227
12.1.2	<i>Level of no motion and “dynamic height”</i>	228
12.2	<i>Coastal upwelling, complete picture</i>	229
12.2.1	<i>Internal dynamics on shelf</i>	230
12.3	<i>Exercises</i>	234
13	<i>Sverdrup balance and basin-scale circulation</i>	236
13.1	<i>Circulation</i>	237
13.2	<i>Conservation of potential vorticity</i>	238
13.3	<i>Sverdrup transport</i>	240
13.3.1	<i>Alternative derivation: Ekman pumping</i>	242
13.4	<i>Return flow</i>	244
13.5	<i>Example problems</i>	245

Foreword

These are the course notes for Descriptive Physical Oceanography, taught at the University of Victoria, School of Earth and Ocean Sciences, a 13-week undergraduate course.

Unlike many physical oceanography courses, students taking this course are not expected to have studied vector calculus or differential equations, nor are they expected to have learned classical mechanics beyond what is typically offered in a first-year physics course. Many of the students have a deep interest in understanding the ocean and often approach the course with considerable intuition. The challenge of a course like this is to offer more than just a list of facts, but rather to try to give a feeling for the forces that drive the ocean, without becoming too bogged down in mathematical formalism. As such, this course tries to strike a balance between knowing about the various physical phenomena in the ocean and understanding the dynamics that give rise to them, while still giving students some practice in practical calculations.

The course is divided into three unequal parts: overturning circulations driven by lateral buoyancy differences and mixing; waves and tides; and wind-driven circulations. We start with estuaries as a nice introduction to overturning circulations, but also because the students do work in Saanich Inlet, and understanding estuaries early is very helpful.

Part I

Overturning circulations

4

Ocean Stratification and the Mixed Layer

The previous chapter discussed the heat and freshwater fluxes that act on the ocean surface, and the turbulent mixing that transports those properties into the interior. In this chapter we look at how these forcings shape the vertical structure of the ocean: a warm, well-mixed surface layer sitting above a strongly stratified thermocline, which in turn decays into weaker but persistent stratification all the way to the sea floor.

4.1 Vertical structure of the ocean

Most of the ocean's variability in temperature and salinity is in the upper ocean (figure 4.1), with weaker gradients at depth. This structure applies through much of the ocean, with most of the variation taking place across a *thermocline* that is on the order of 1000 m thick. Below the thermocline, temperature and salinity gradients are much weaker, but they do not vanish — the deep ocean retains stable stratification all the way to the bottom.

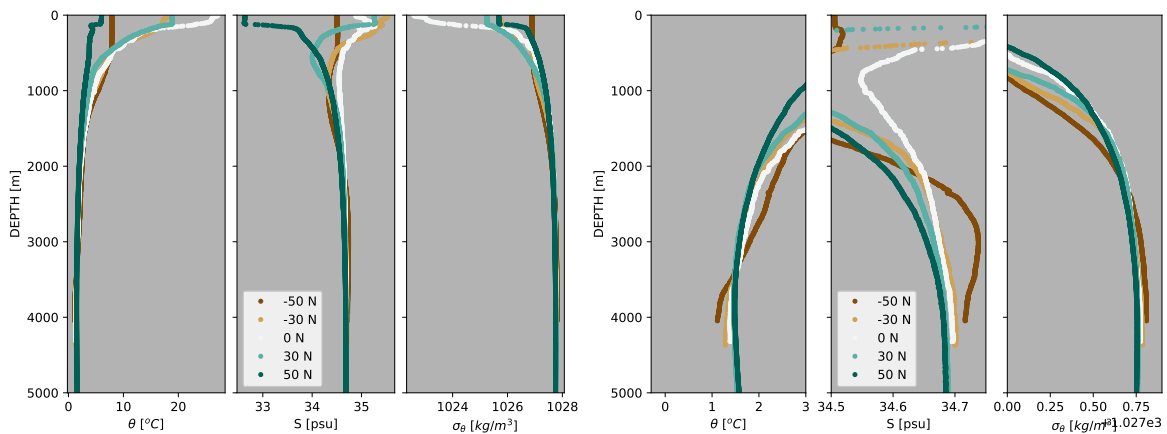


Figure 4.1: Example θ and S profiles from the Pacific Ocean at five different latitudes. The two plots on the right are zoomed versions of the plots on the left.

Looking at just the linear-axis profiles on the left of figure 4.1, you

might think the deep ocean is uniform. But the zoomed profiles on the right show that temperature and salinity continue to change with depth even below 2000 m. In many parts of the ocean, the decay of stratification with depth looks roughly exponential, which means that a logarithmic scale reveals structure that a linear plot hides. This is worth keeping in mind whenever you look at ocean profiles: the deep gradients are small, but they are real and they matter for the circulation.

4.2 Buoyancy frequency

To quantify stratification, we define the *buoyancy frequency* (also called the *Brunt-Väisälä frequency*), N :

$$N^2 = -\frac{g}{\rho_0} \frac{d\rho}{dz} \quad (4.1)$$

where g is gravitational acceleration, ρ_0 is a reference density (approximately 1025 kg m^{-3}), and $d\rho/dz$ is the vertical density gradient. Because density generally increases downwards (and z points up), $d\rho/dz$ is typically negative and N^2 is positive. N has units of rad s^{-1} , and is often reported in cycles per hour (cph).

Physically, N is the frequency at which a parcel of water would oscillate if displaced vertically in a stably stratified fluid. If you push a parcel of water downward into denser surroundings, buoyancy pushes it back up; it overshoots, finds itself lighter than its surroundings, and sinks back. The stronger the stratification, the higher the frequency of this oscillation, and the harder it is for turbulence to overturn the water column.

Typical values of N in the ocean span about two orders of magnitude:

- In the thermocline: $N \sim 10^{-2} \text{ s}^{-1}$ (about 6 cph), corresponding to oscillation periods of about 10 minutes.
- In the deep ocean: $N \sim 10^{-3} \text{ s}^{-1}$ (about 0.6 cph), with periods closer to two hours.
- In the mixed layer: $N \approx 0$, because the water is nearly uniform in density.

Box 4.2.1: Estimating N^2 from a profile

Suppose a CTD profile shows $\rho = 1026.5 \text{ kg m}^{-3}$ at 500 m depth and $\rho = 1027.0 \text{ kg m}^{-3}$ at 1000 m depth. Then:

$$N^2 = -\frac{g}{\rho_0} \frac{\Delta \rho}{\Delta z} = -\frac{9.8}{1025} \frac{(1027.0 - 1026.5)}{(500 - 1000)} = -\frac{9.8}{1025} \frac{0.5}{-500} \approx 1 \times 10^{-5} \text{ s}^{-2}$$

so $N \approx 3 \times 10^{-3} \text{ s}^{-1}$, or about 2 cph. This is a typical mid-depth value.

You will calculate N^2 from your own CTD data — it is one of the most useful quantities to extract from a density profile.

4.3 OvertURNS

The ocean is “stably stratified” ($N^2 > 0$), with denser water sitting below lighter water almost always. An *overtURN* occurs when this is violated: a patch of water where denser water has been lifted above lighter water by turbulence. These are the fingerprints of active mixing in the ocean interior.

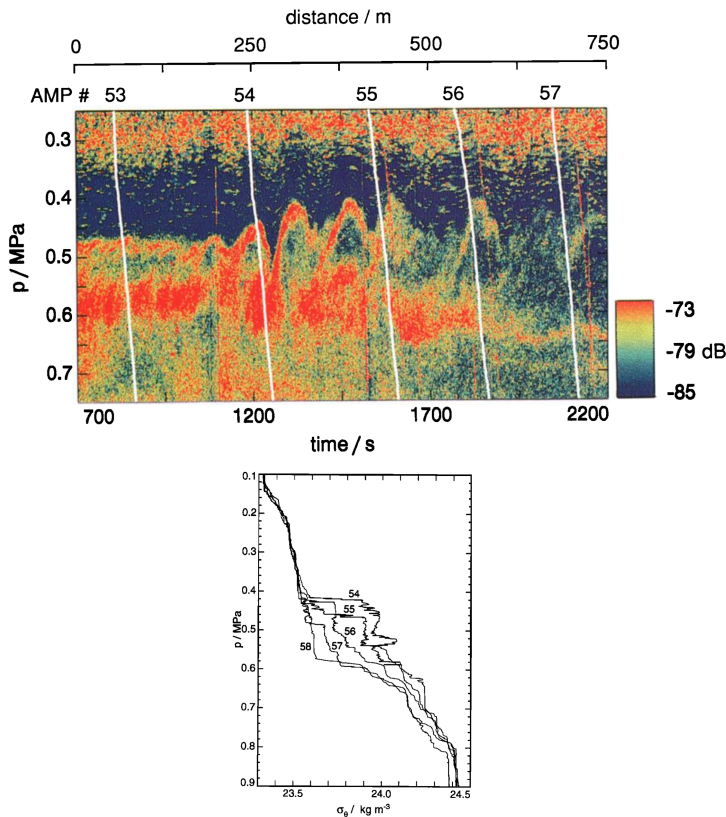


Figure 4.2: Example of a turbulent region in Admiralty Inlet (near Seattle, WA) where the density profile is overturned. Flow is from left to right and the turbulence appear as “waves” or “braids” in the acoustic backscatter (color); the (estimated) path of an instrument is shown in white. Right hand plot shows the density profiles from the instrument, and clearly density becomes more disordered downstream of the turbulence [?].

Overturns are relatively rare. In a typical deep-ocean profile, less than a few percent of the water column is actively overturning at any given time. In the thermocline, where stratification is strong, overturns are smaller still — typically just a few meters tall. In weakly stratified deep water, they can be somewhat larger, but even there, most of the water column is stably stratified at any instant.

This tells us something important: the interior ocean is a very quiet place. Turbulent mixing happens in brief, localized bursts, not as a steady background churn. Most of the mixing energy comes from a few specific sources — internal waves breaking, flow over rough topography, and shear instabilities — and these are concentrated in particular locations. We will return to where this mixing energy comes from when we discuss tides and internal waves.

Q: If the ocean is so rarely overturning, how can mixing be important for the large-scale circulation?

4.4 The surface mixed layer

The upper ocean is usually topped by a *mixed layer*, where the action of waves and wind have combined to mix the water uniformly. This can be seen in all the example profiles in figure 4.3, except for the profile at the equator (white) as a region of homogeneous water in both the temperature and salinity profiles. At 50°S, this layer is particularly thick, extending over 150 m deep. In the plot shown it looks like it is much deeper, but there is some slight temperature and salinity difference with depth that stops the water from being completely homogeneous.

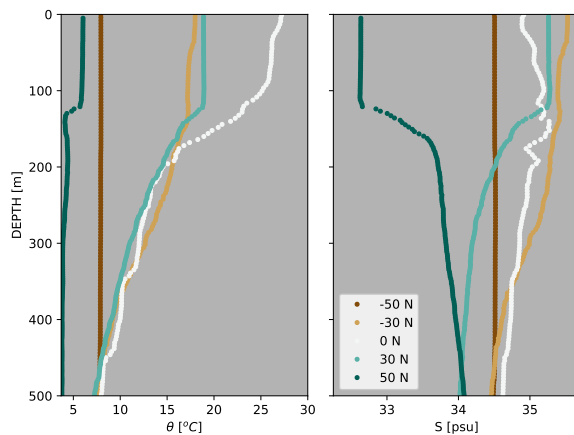


Figure 4.3: Example θ and S profiles from the Pacific Ocean at five different latitudes. The two plots on the right are zoomed versions of the plots in figure 4.1.

In practice, the mixed layer depth is usually defined using a density threshold: the depth at which density has increased by some small amount (typically $\Delta\rho = 0.03 \text{ kg m}^{-3}$) relative to the surface. This is more robust than using temperature alone, since salinity also affects density and can create shallow stratification even when the

temperature profile looks uniform.

The dynamics of the mixed layer are very important given that they mediate the flux of temperature and gasses between the ocean and the atmosphere. There are three things that deepen the mixed layer:

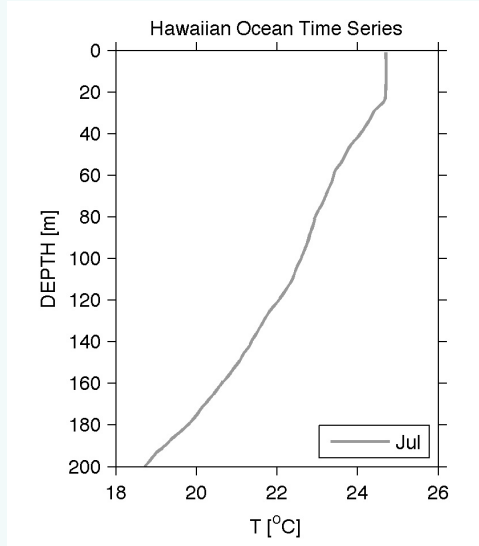
1. Cooling from evaporation and sensible heat loss
2. Mixing from wind-driven turbulence and waves
3. Shear-driven turbulence at the base of the mixed layer.

The first term is relatively straight forward: when the water cools, it is denser than the water below, creates an overturn and quickly goes unstable and mixes. The wind-driven term is hard to parameterize, but accords with our usual understanding of waves as turbulent forces. The last term is because the wind pushes the mixed layer, and hence it develops momentum that is different in speed and direction from the water below. This generates a *shear* and these can go unstable (chapter 3).

The mixed layer will *restratify* due to

1. Heating or freshwater
2. Lateral advection of more buoyant water (often from equatorward).

Box 4.4.1: Heat budget of the mixed layer



Consider the temperature profile above from the HOT Station north of Hawaii. In the summer, the mixed layer is warm and shallow, at roughly 25°C and perhaps 50 m deep. Below this is the permanent thermocline. Over the winter, cooling will deepen the mixed layer. Suppose the mixed layer cools and deepens until the temperature is uniform at about 24°C to a depth of 100 m.

We can estimate the heat lost during this process. The heat content change per unit area of ocean surface is:

$$\Delta Q = \rho c_p \int_0^H \Delta T(z) dz$$

where $c_p \approx 4000 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of seawater and $H = 100 \text{ m}$ is the final mixed layer depth. Estimating from the profile, the average temperature change over the upper 100 m is roughly 1°C, so:

$$\Delta Q \approx 1025 \times 4000 \times 1 \times 100 \approx 4 \times 10^8 \text{ J m}^{-2}$$

If this cooling happens over approximately 6 months ($\approx 1.5 \times 10^7 \text{ s}$), the average net heat flux is:

$$\bar{Q}_{net} = \frac{\Delta Q}{\Delta t} \approx \frac{4 \times 10^8}{1.5 \times 10^7} \approx 25 \text{ W m}^{-2}$$

This is a modest net heat loss, well within the range of observed net surface heat fluxes, confirming that the seasonal cycle of the mixed layer is consistent with what we know about surface forcing.

4.4.1 Seasonal and diurnal cycles

This cycle leads to an annual/seasonal cycle of the upper ocean mixed layer (figure 4.4), with cooling in the winter, and warming in the summer. In this case, the mixed layer gets down to about 100 m by later spring, and then starts shoaling again in the summer. This often leads to a situation where there are two thermoclines in the summer; the *permanent thermocline* deeper than 100 m, and a *seasonal thermocline* beneath the summer mixed layer. Note that in this example, the mixed layer is also strongly influenced by salinity, so the concept of the *halocline* is important to understand the annual cycle and why the mixed layer only penetrates to about 100 m.

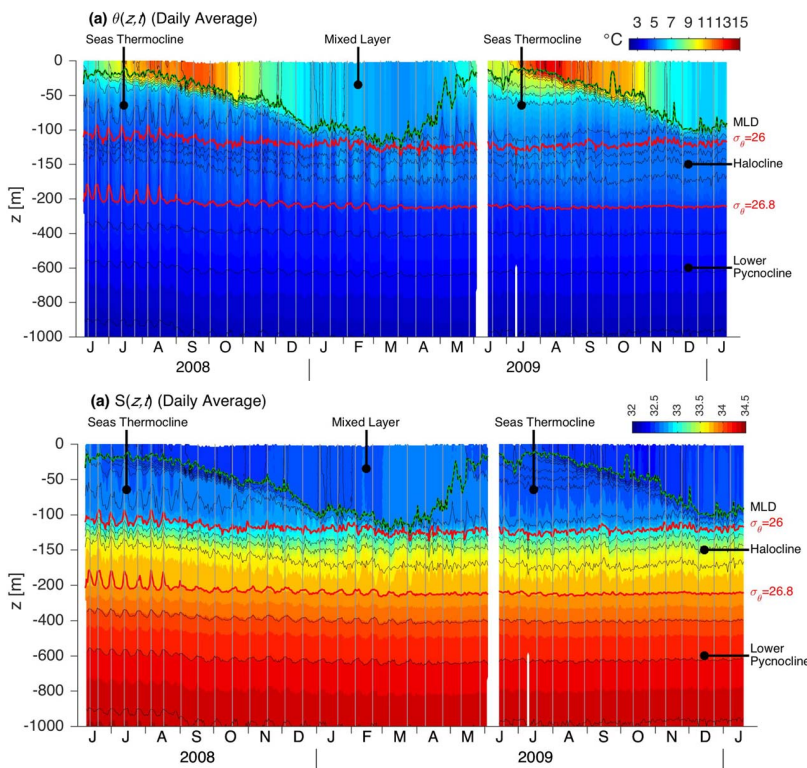


Figure 4.4: Temperature and Salinity in the upper ocean over an annual cycle at Station Papa (Northeast Pacific Ocean: 50 N, 145 W) [?]

The mixed layer also evolves on a daily time scale, with warming in the day and cooling at night, of course modified by the wind conditions (figure 4.5). In this example when the ocean is being warmed by the atmosphere, the mixed layer becomes very shallow (figure 4.5B) and turbulence deeper in the water column is suppressed (figure 4.5C). At night the water column cools and turbulence gets to deeper depths.

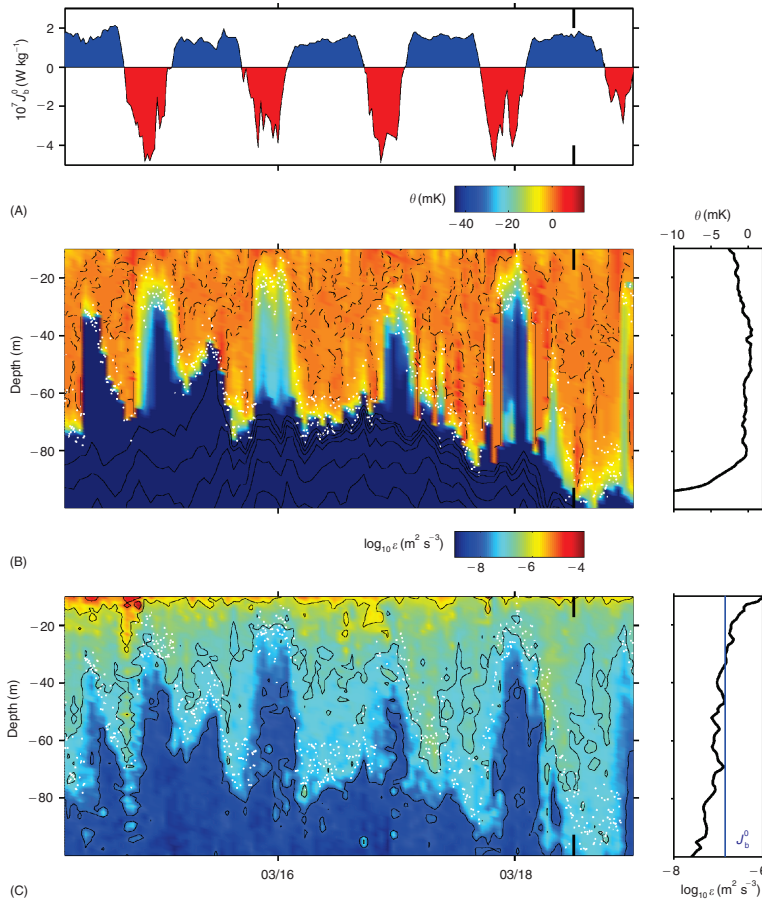


Figure 4.5: Mixing in the upper ocean in NE Pacific summer. A) heat flux, red being ocean heating, blue cooling. B) Temperature anomaly relative to depth-mean. So this shows the vertical variation, but not the cooling and heating with time. White dots indicate the base of the mixed layer. C) Turbulence dissipation rate.

4.4.2 Global mixed layer depths

Given the variety of forcings it is not surprising that the ocean mixed layer has a wide variety of depths through the worlds oceans (figure 4.6). Depths are greatest in the winter, and reach quite great depths near the poles. Depths also tend to be largest on the windward side of the basins, where there are the greatest heat fluxes.

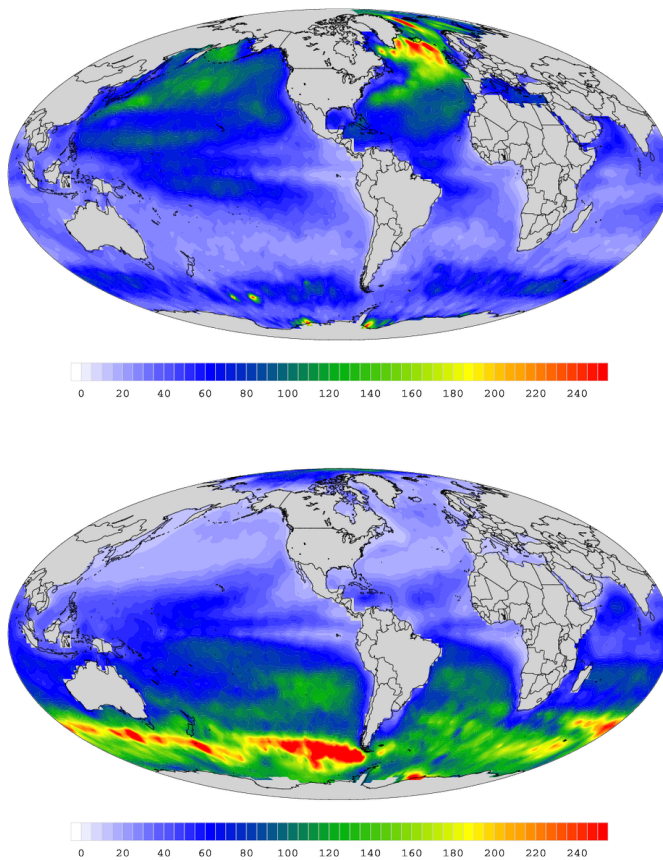
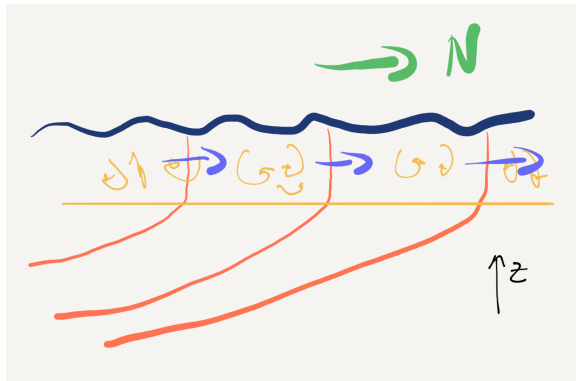


Figure 4.6: Mixed layer depth in the boreal and austral winters.

4.5 *Restratification of the mixed layer*



A north-south section of the ocean in the northern hemisphere will tend to look like the schematic above. If there were no mixing in the upper ocean, the orange isopycnals would tend to flatten, driving a flow to the North. This happens all the time in the upper ocean where there is a warmer (lighter) mixed layer equatorward of a point. Hence heat is being constantly fluxed poleward by this restratification, and if the mixed layer is to stay mixed, it must mix via turbulence and lose heat to the atmosphere. So, when the north Pacific warms in the summer, it is not only the result of the local warming by the sun and lack of heat losses due to evaporation, but is also due to a northwards transport of heat by this restratification.

4.6 *What maintains the deep stratification?*

The ocean below the mixed layer is stably stratified, with dense cold water at the bottom and lighter water above. This stratification is maintained by a balance between the sinking of cold dense water at high latitudes (which we will discuss in chapter 6) and the slow upward diffusion of heat from above. This is exactly the same process that happens in an estuary and drives the estuarine circulation. The details of this balance — how much mixing is required, and where it occurs — remain an active area of research.

4.7 *Exercises*

Interpret the evolution of the surface mixed layer Based on the following plot, try and answer the following questions:

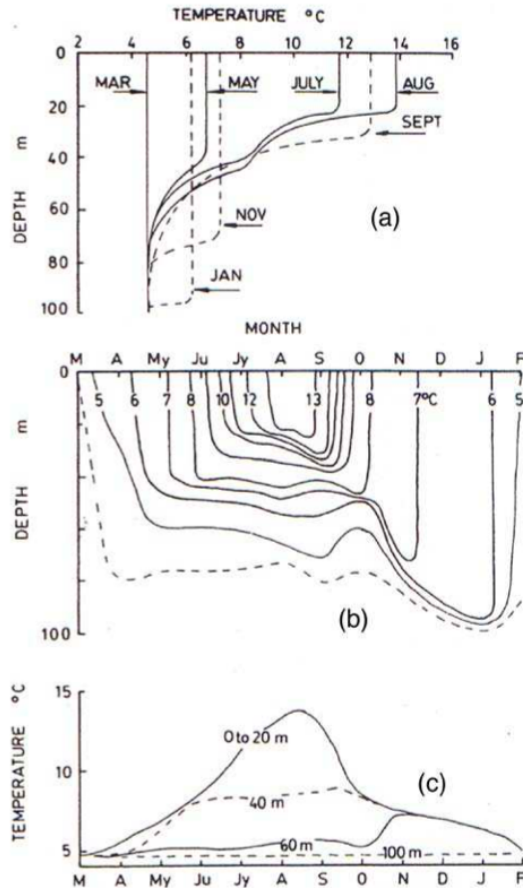


Fig. 4.7. Growth and decay of the seasonal thermocline at 50°N, 145°W in the eastern North Pacific as (a) vertical temperature profiles, (b) time series of isothermal contours, and (c) a time series of temperatures at depths shown.

From Emery et. al. 2007

- Why does the mixed layer deepen in the winter?
- Why does it shoal in the spring and summer?
- Why do you think there is often a spring bloom at Station Papa?
Why does it stop?

Buoyancy frequency Consider the density profile below from a CTD cast. Density increases approximately linearly from $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ at 200 m to $\sigma_\theta = 27.0 \text{ kg m}^{-3}$ at 1000 m.

- Estimate N^2 and N for this depth range.
- What is the period of oscillation of a displaced water parcel?

- If a turbulent overturn in this region were 5 m tall, roughly how long would it take buoyancy to collapse it? (Hint: the buoyancy time scale is $1/N$.)
- Why are overturns in the thermocline generally much smaller than overturns in the deep ocean?

Part II

Waves and Tides

Part III

Wind-driven circulations