The ocean and its circulation

First, some numbers about the ocean. Seventy percent of the earth's surface is covered by the ocean, which has mean depth of ~4km. Recall that the atmosphere has a heat capacity corresponding to a ~3 meter thick water layer, and note the huge heat capacity of the ocean in comparison. We have noted that the large heat capacity of the ocean reduces the diurnal seasonal cycle. The ocean also has a low albedo (~0.1) and is the primary source of moisture. It is also a major player in biogeochemical cycles.

Like the atmosphere, the ocean is also a stratified fluid on the rotation Earth, but there are some important differences:
1. Different equation of state: water is almost incompressible; there is no source of latent heat in the ocean but salinity affects the density of seawater
2. Oceans are in general laterally bounded by continents. This has some important consequences on the dynamics.
3. The ocean is forced at the top.

Seawater is salty. Salt and temperature are the two main factors affecting the density. Density also depends on pressure. But for most of our discussion, it's not as important.

The equation of state for seawater is determined from experiments. Salinity of sea water is determined from its conductivity (salty water has higher conductivity) and reported in "Practical Salinity Unit" or psu, which is nearly equal to g/kg of salt in sea water. Note that fresh water is most dense at 4C. Therefore, as a fresh water lake is cooled from above, it will convect when temperature is above 4C, but will become stably stratified when surface temperature drops below 4C. This is of course why ice forms on the top in freshwater lakes. This is an important fact as ice, being a solid, is a good insulator that reduces further loss of heat from the ocean/lake to the atmosphere. Otherwise, freshwater lakes would be frozen solid. For seawater with salinity > ~25psu (the regime that the ocean resides), however, density decreases monotonically with temperature. This implies that if salinity is well mixed, as surface water is cooled, it will sink to the bottom and ice will form only when the whole column is cooled to the freezing point, i.e. either there is no ice or the whole column is frozen.

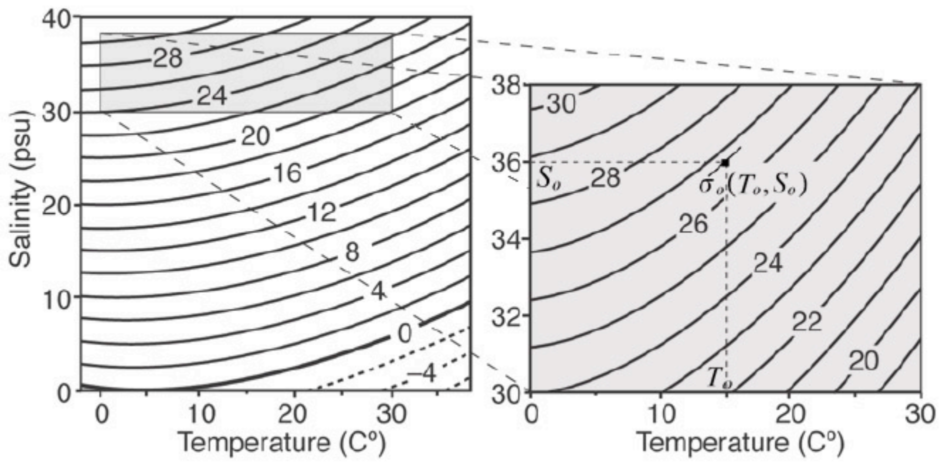


Figure 9.2: Contours of seawater density anomalies ($\sigma = \rho - \rho_{ref}$ in $\mathrm{kg\,m^{-3}}$) plotted against salinity (in psu $= \mathrm{g\,kg^{-1}}$) and temperature ($^\circ$C) at the sea surface. Note that sea water in the open ocean has σ in the range 20 to 29 $\mathrm{kg\,m^{-3}}$, T in the range $0 - 30\,^\circ\mathrm{C}$ and S in the range $33 - 36$ psu. The panel on the right zooms in on the region of oceanographic relevance. An approximation to the equation of state in the vicinity of the point $\sigma_{\boldsymbol{o}}(T_{\boldsymbol{o}},\, S_{\boldsymbol{o}})$ is given by Eq.(9.5).

Within the temperature and salinity range seen in the open ocean, salty water is denser than fresh water and cold water is denser than warm water. The effect of salinity on density is relatively uniform while the effect of temperature is greater at higher temperatures (and also at higher pressures). Some examples are given below, where we show the thermal expansion coefficient $\alpha_T = -\dfrac{1}{\rho_{ref}}\dfrac{\partial \rho}{\partial T}$ and the dependence of density on salinity

$$\beta_S = \frac{1}{\rho_{ref}}\frac{\partial \rho}{\partial S}$$

Density is shown as departure from a reference density 1000kg/m^3.

Surface			
T_o (°C)	-1.5	5	15
α_T $(\times 10^{-4}\,\mathrm{K}^{-1})$	0.3	1	2
S_o (psu)	34	36	38
β_S $(\times 10^{-4}\mathrm{psu}^{-1})$	7.8	7.8	7.6
σ_o $(\mathrm{kg\,m^{-3}})$	28	29	28
Depth of 1 km			
T_o (°C)	-1.5	3	13
α_T $(\times 10^{-4}\,\mathrm{K}^{-1})$	0.65	1.1	2.2
S_o (psu)	34	35	38
β_S $(\times 10^{-4}\mathrm{psu}^{-1})$	7.1	7.7	7.4
σ_o $(\mathrm{kg\,m^{-3}})$	-3	0.6	6.9

Table 9.4: The dependence of σ, α_T and β_S on T and S at two levels in the ocean, at the surface and a depth of 1km.

Temperature and salinity distributions

Distributions of ocean surface temperature can be see following the link on the course webpage to the Remote Sensing system website, and click on optimally interpolated sea surface temperature (SST) http://www.ssmi.com/sst/sst_data_daily.html?sat=mw_ir. There are a few things to note: the east/west asymmetry; the seasonal migration of SST; the difference between the two equinoxes. You can also zoom onto different regions and see the Gulf Stream in the Atlantic and the Kuroshio Current in the Pacific, and the strong meridional gradients in the southern ocean.

Unlike SST, there are no salinity measurements from space, so surface salinity maps are based on measurements by ships that go around the globe. The main features here are that the subtropical oceans are saltier than the tropics and midlatitudes (why?); the Atlantic is also saltier than the Pacific.

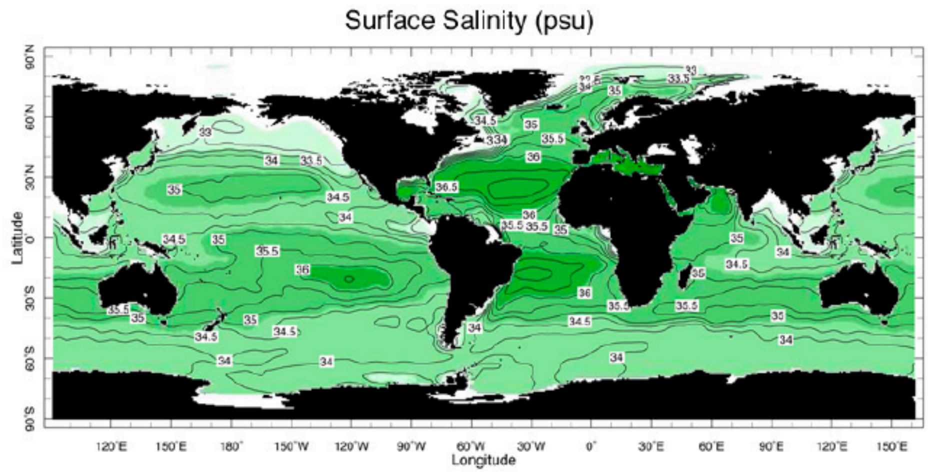


Figure 9.4: The annual mean salinity distribution at the surface of the ocean in psu. Darker green represents salty fluid. Data from Levitus World Ocean Atlas, 1994.

Now here are some cross sections of zonal-average temperature and salinity. Note that most of the vertical gradients are concentrated in the upper 1km. This is the thermocline (excluding the mixed layer where temperature/salinity is uniform with depth), which is ~600m in subtropics but shoals towards the tropics. The warmer water at low latitudes and higher salinity in the subtropics are expected. Note the fresher water going from the surface of the southern ocean to about 1km at lower latitudes. This is the Antarctic intermediate water. Also note that water is colder near the surface than at 1km depth in the southern ocean. How could that be? A similar situation occurs in the Arctic Ocean.

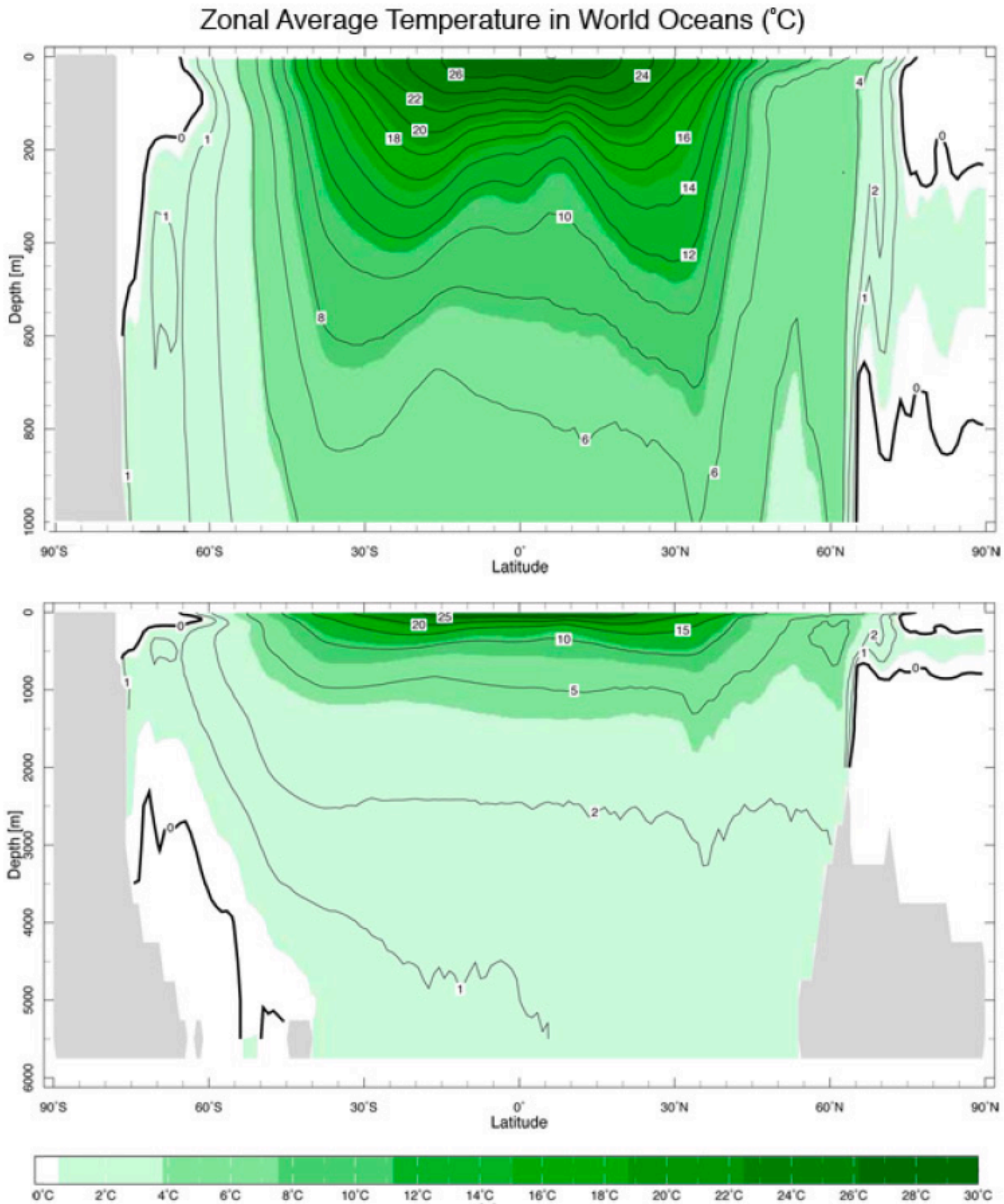


Figure 9.5: Annual-mean cross-section of zonal-average potential temperature (in °C) in the world's oceans: top shows upper 1 km, bottom shows the whole water column. Dark shading represents warm water. Note the variable contour interval in the bottom plot. Data from the Levitus World Ocean Atlas 1994.

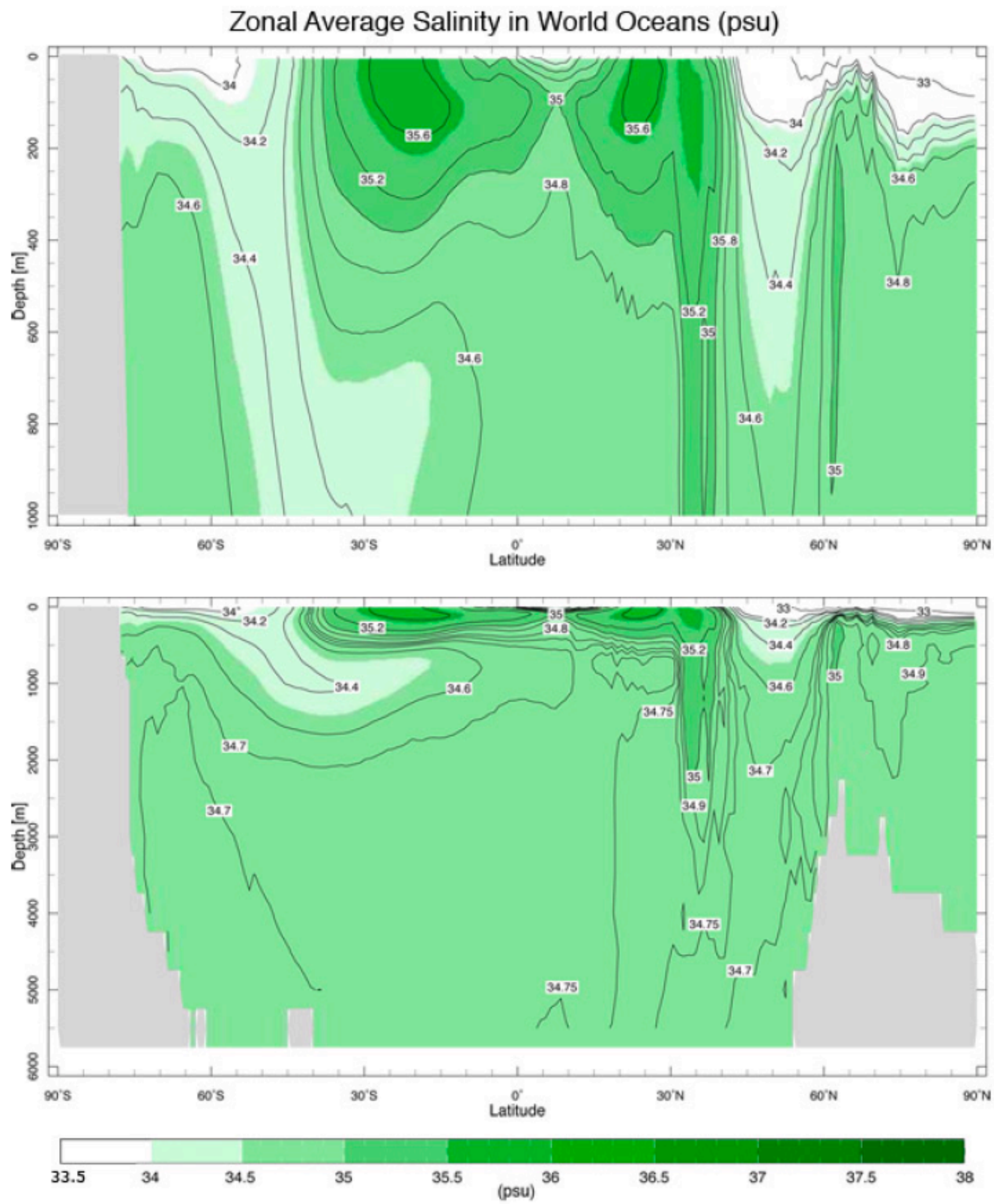


Figure 9.6: Annual-mean cross-section of zonal-average salinity (in psu) in the world's oceans: top shows upper km, bottom shows the whole water column. Darker represents salty water. Data from the Levitus World Ocean Atlas 1994.

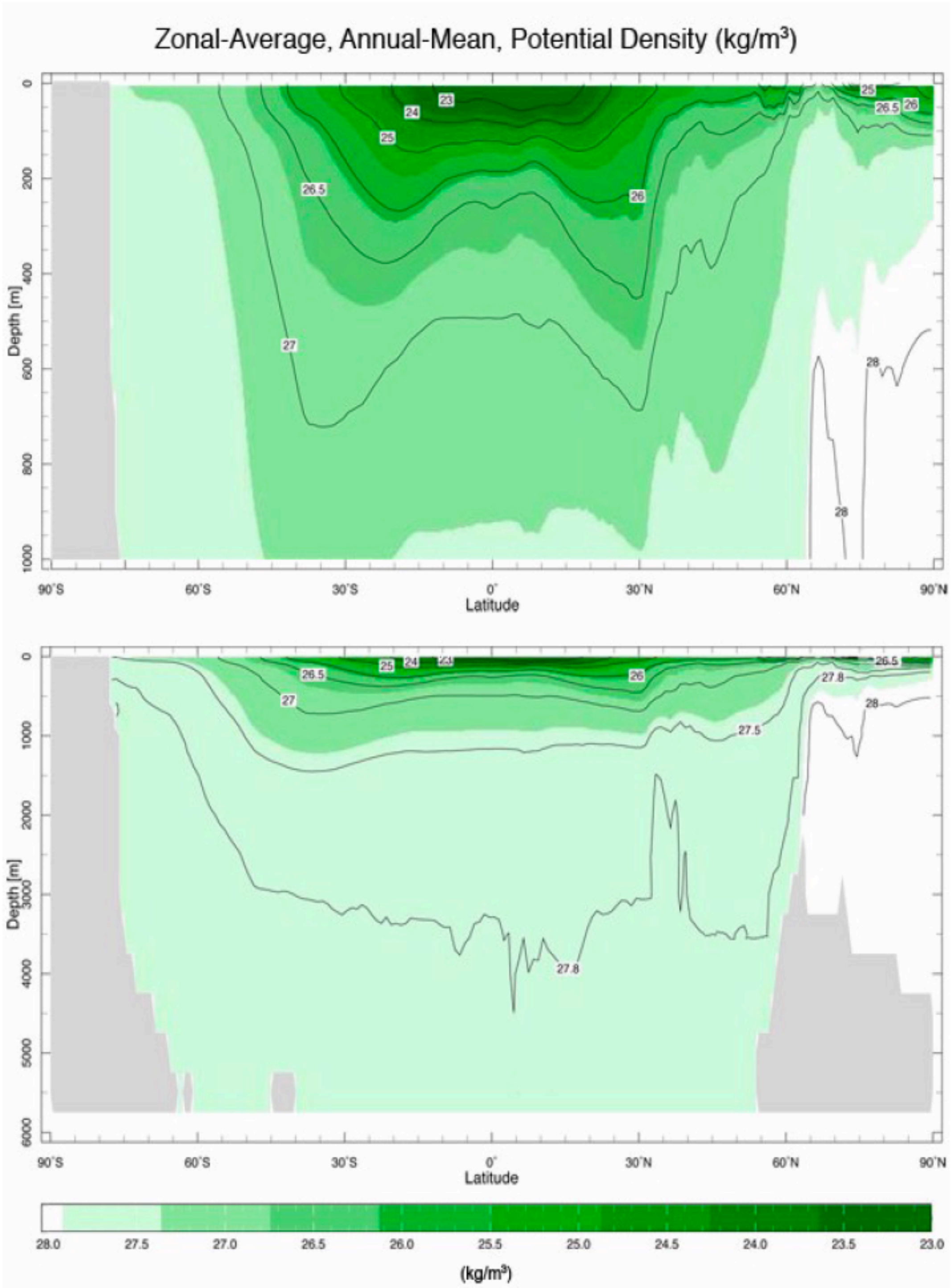


Figure 9.7: Annual-mean cross-section of zonal average potential density anomaly $\sigma = \rho - \rho_{ref}$ (in $\mathrm{kg\,m^{-3}}$) for the world oceans (referenced to the surface): top shows upper 1 km, bottom shows the whole water column. Note that the contour interval is not uniform. Data from the Levitus World Ocean Atlas 1994.

The answer is that the surface tends to have lower salinities. This is due mostly to greater precipitation (compared to evaporation) over these regions, although in the Arctic Ocean, the supply of freshwater from rivers from the surrounding continents is an important contributor. This fact is important because it allows cold water to stay at the surface and form sea ice. It has been hypothesized that if one diverts river flows away from the Arctic Ocean, it might lead the Arctic to be either completely ice free or completely frozen from the surface to bottom (!).

In the potential density (density when brought to a reference pressure) plot, note that if a parcel conserves its potential density then water in the deep ocean should link with surface water at high latitudes.

Below is a cross section of T and S (salinity) as a function of depth in the Atlantic from Conductivity-temperature-depth (CTD) measurements from a ship (these are expensive measurements to make!). You can see the thermocline quite clearly.

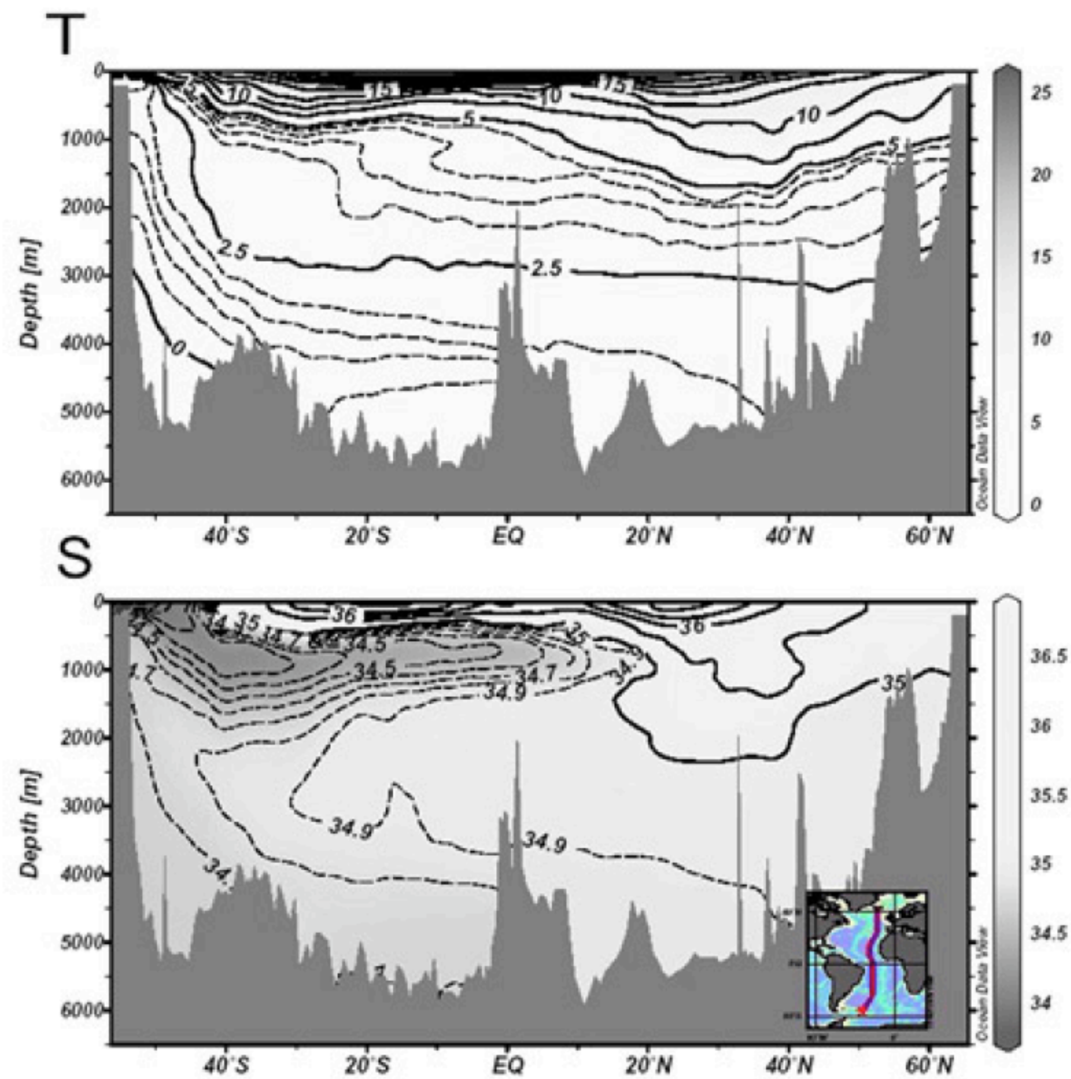


Figure 9.9: Hydrographic section along (roughly) 25°W through the Atlantic Ocean — see inset. Top: Potential temperature contoured every 2.5 °C (solid) and every 0.5 °C (dotted). Bottom: Salinity in psu. Values greater than 35 are contored every 0.5psu (solid): values less than 35 are plotted every 0.1 psu (dotted). Figure produced using Ocean Data View.

The mixed layer is a layer at the surface of the ocean that is well mixed due to stirring by winds and convection. Processes that affect the mixing are illustrated below. Sunlight can penetrates 10m or more into clean water. So depend on how clean the water is, solar heating can be distributed over a few meters from the surface. This has a stabilizing effect. Longwave flux, surface sensible and latent heat fluxes are in general directed out of the ocean, and have a destabilizing effect. Wind can also generate turbulence. Rainfall, on the other hand, can create a layer of fresh water and exert a stabilizing effect. The depth of the mixed layer varies depending on the relative strength of the above factors, and is typically 50-100m deep. It's shallower in sunny, warm, calm days, and deeper in cold, and stormy days. It's particularly deep in the North Atlantic and the Southern ocean, where oceanic deep convection takes place.

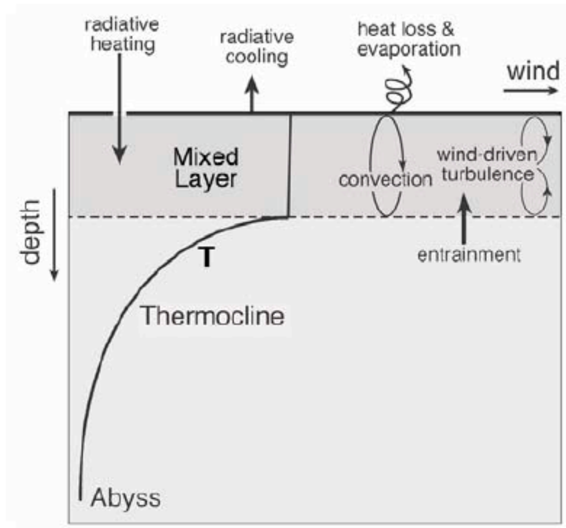
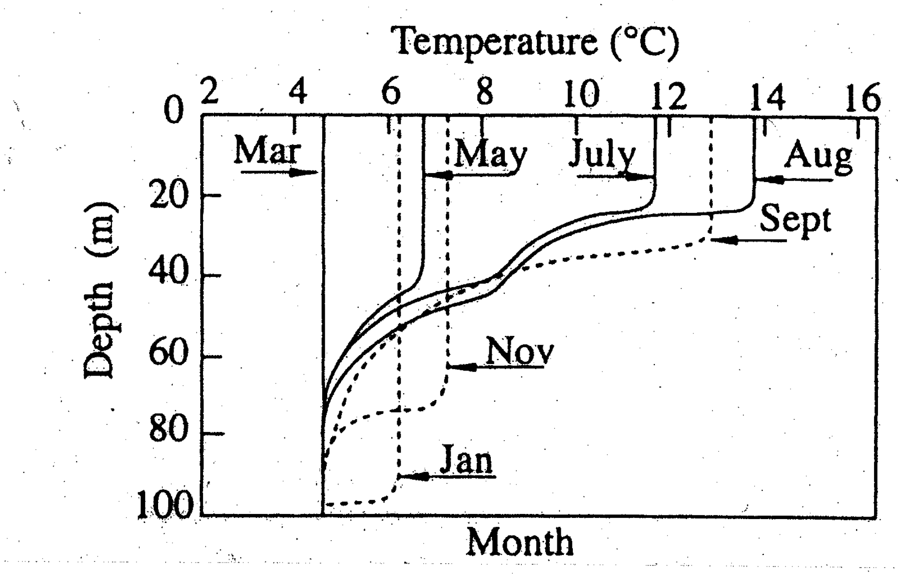


Figure 9.11: A schematic diagram showing processes at work in the mixed layer of the ocean. Note that the vertical scale of the mixed layer relative to the thermocline is greatly exaggerated.



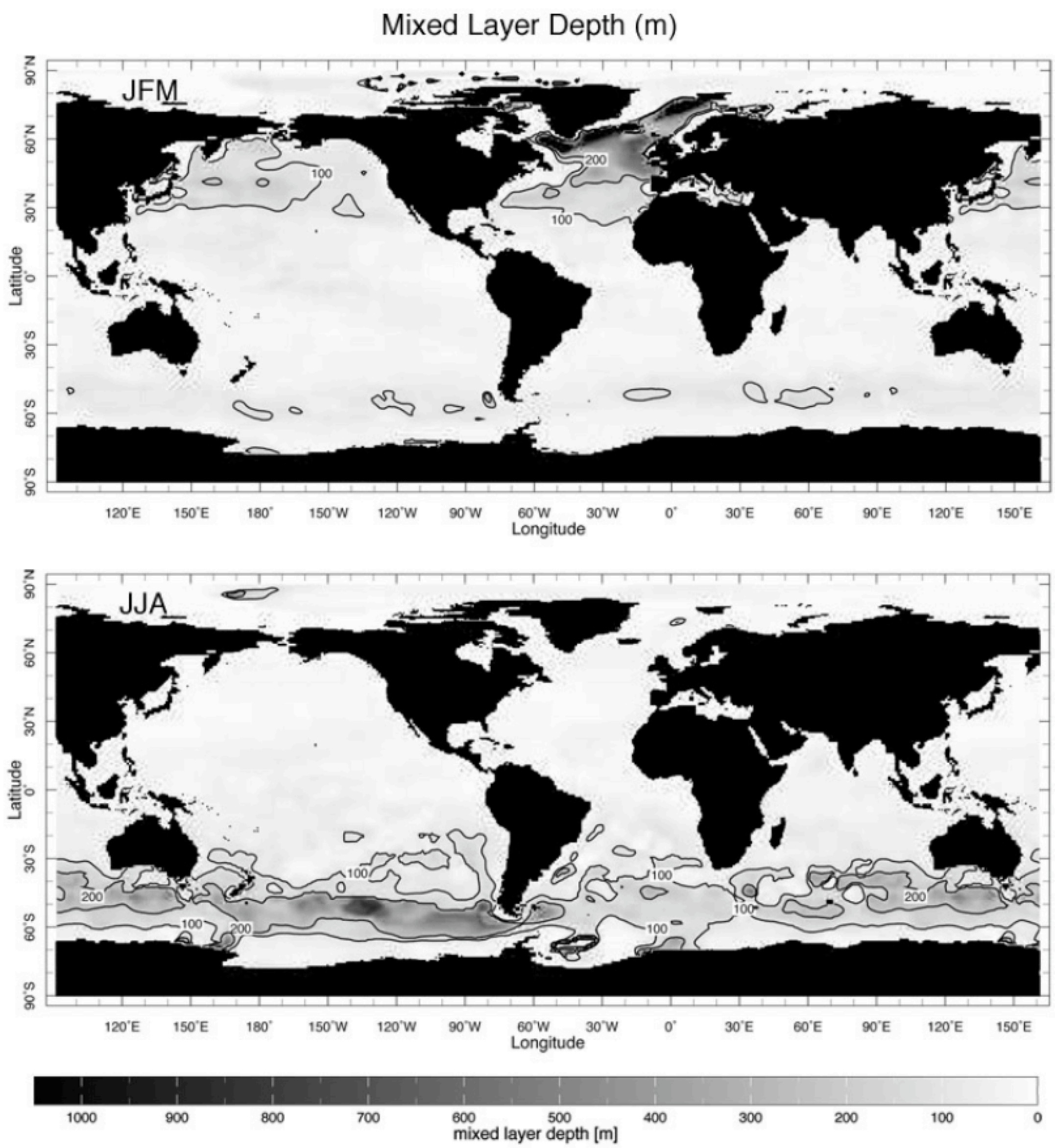


Figure 9.10: Mixed layer depth (in m) in (top) JFM (January, February, March, Northern hemisphere winter) and (bottom) JJA (June, July, August, Southern hemisphere winter). Black contours mark the 100 and 200 m mixed layer depth isopleths. Data from the Levitus World Ocean Atlas 1994.

The mixed layer readily responds to meteorological forcing on daily and seasonal timescales, while the interior of the ocean (i.e. below the mixed layer) responds on interannual, decadal, and longer timescales.

The ocean, being a stably stratified fluid (below the mixed layer), also supports gravity waves. In the thermocline, the buoyancy period is roughly 20min. When these waves break, they incur mixing, which turns out to be important for the circulation in the abyss or the thermohaline circulation.

The observed mean circulation

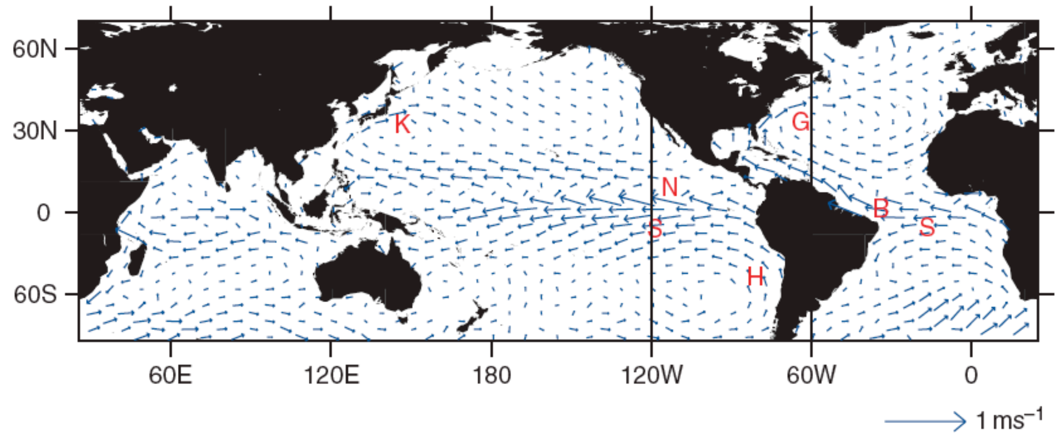


Figure: Major surface currents. Note the Humboldt Current (labeled H) is often called the Peru Current.

Despite the complicated nature of the currents, we see in general anticyclonic subtropical gyres and cyclonic subpolar gyres. At the western edges of the ocean basins, the currents tend to be strong. The Kuroshio and the Gulf Stream can have current speeds > 100cm/s. (They don't show up well in the above plot because of the coarse resolution, but are a lot clearer in Figs. 9.14, 9.15 in Marshall and Plumb.) There are no intense currents on the eastern boundaries of ocean gyres.

In the tropics are the strong westward North and South Equatorial currents, and the eastward Equatorial Current in between (just north of the equator). There is also an Equatorial Undercurrent. These are best seen at the TOA array webpage http://www.pmel.noaa.gov/tao/.

In the southern ocean, there is a strong zonal flow called the Antarctic Circumpolar Current (ACC).

With a speed of a few tens of cm/s, the equatorial currents can cross Pacific in 2 years. It takes roughly this much time for the ACC to go around Antarctica. It takes longer (~5years) to go around the subtropical gyre in the Atlantic Ocean.

These currents extend into the interior of the ocean (i.e. below the mixed layer) but become weaker at depth.

The ocean obeys the same fluid dynamics as the atmosphere, except here we can simplify things by considering the fluid being incompressible.

$$\frac{Du}{Dt} + \frac{1}{\rho_{ref}}\frac{\partial p}{\partial x} - fv = \mathcal{F}_x ;$$
$$\frac{Dv}{Dt} + \frac{1}{\rho_{ref}}\frac{\partial p}{\partial y} + fu = \mathcal{F}_y ;$$

F here can be from interior friction, boundary friction or surface stress.
The relative importance of the D(u,v)/Dt terms compared to the Coriolis term is, as we have learned, measured by the Rossby number. For a midlatitude ocean gyre

$$R_o = \frac{U}{fL} = \frac{10cm/s}{10^{-4}/s2000km} \sim 10^{-3}$$

Indeed, currents away from the boundaries, the equator, and the surface are approximately in geostrophic balance.

Pressure in the ocean can be calculated from hydrostatic balance:

$$\frac{\partial p}{\partial z} = -g\left(\rho_{ref} + \sigma\right)$$

At depth -z relative to the geoid, the pressure is

$$p(z) = p_s - g\rho_{ref}(z-\eta) - \int_{\eta}^{z} g\sigma dz'$$

Ps is the surface atmosphere pressure, and η is the free surface height. The horizontal pressure gradients are

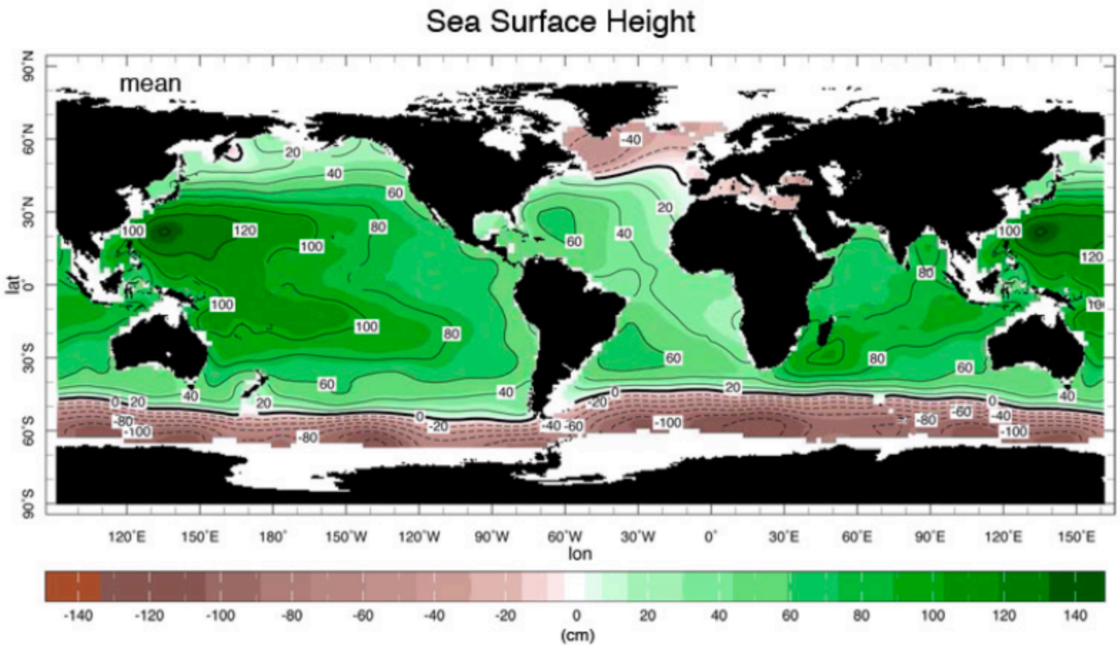
$$\nabla p = \nabla p_s + g\rho_{ref}\nabla\eta - \nabla\int_{\eta}^{z} g\sigma dz'$$

Effect from Ps is in general small. The atmospheric winds that are roughly in geostrophic balance with the surface pressure are typically 10m/s. Take into account the density difference between air and water, the same pressure gradient can balance ocean currents of 1cm/s, which is smaller than the currents observed on a basin scale. Let's also for the moment neglect horizontal density difference (the last term in the above equation). This is fine very near the surface (say, the top 100 meters). Very near the surface, it's not really in geostrophic balance as the effect of surface stress must be considered. But the magnitude of the force from surface stress is in general smaller compared to the pressure gradient force and the Coriolis force, so we can still proceed assuming approximate geostrophic balance.

What this implies is that when there is a strong surface current, there should be a variation in the free surface height. In particular, the ocean surface should be higher to the right of the Gulf Stream than to the left. For a current speed of 10cm/s, we estimate that

$$|\nabla\eta| = |u| f / g = \frac{0.1m/s10^{-4}s^{-1}}{10ms^{-2}} = 10^{-6}$$

or roughly 1 meters in 1000km. This can be readily seen in altimetry measurements of sea surface height.



At depth ($|z|>>\eta$), we can no longer neglect horizontal density gradient. Assume horizontal density gradient is constant with height (or define a vertically averaged density gradient), we have

$$\nabla p = g\rho_{ref}\nabla\eta + g(\eta - z)\nabla\sigma$$

It is observed that in the abyss, ocean currents are weak; the ocean floor has high topography, which slows the currents. What this implies is that the horizontal density gradient must cancel to a large extent the effect of the free surface height variations. From the above equation, we see that in the subtropics where the free surface is high, water needs to be lighter in order to have a small horizontal pressure gradient in the abyss. This can also be seen from the thermal wind relation. For a given stratification, what is the slope of the "isopycnals" (constant density surfaces) that is needed so that a given slope of the free surface is balanced at a depth of H? The slope of the isopycnal surface is

$$\frac{|\nabla\sigma|}{d\sigma/dz} = \frac{|\nabla\eta|\rho_{ref}}{Hd\sigma/dz} = \frac{g|\nabla\eta|}{HN^2}$$

where N is the buoyancy frequency. Plug in reasonable values for N and 1km for H, we find that the slope of the isopycnals is 400 times larger than (and opposite in sign) that of the free surface.

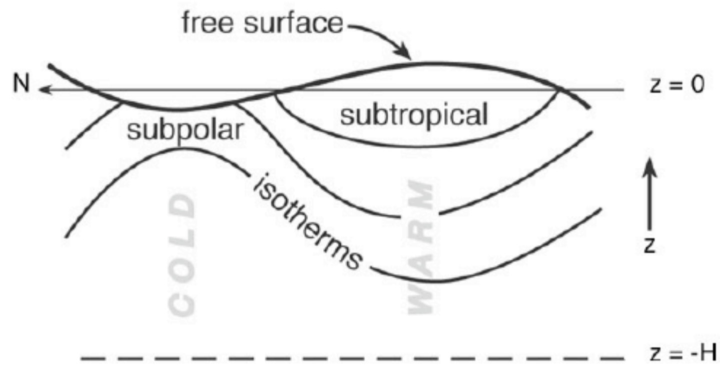


Figure 9.20: Warm subtropical columns of fluid expand relative to colder polar columns. Thus the sea surface (measured relative to the geoid) is higher, by about 1 m, in the subtropics than the pole, and is thus greatly exaggerated in this schematic. Pressure gradients associated with the sea-surface tilt are largely compensated by vertical thermocline undulations, of about 400 m, ensuring that abyssal pressure gradients are much weaker than those at the surface.

This is analogous to the root of mountains (or icebergs); under mountains there are thickened crusts, which are buoyancy, and they cancel out the horizontal pressure gradient created by the orography.
Water in the subtropical ocean is indeed found to be warmer. It's also saltier as we have seen before. Use the equation of state for seawater, we find that the temperature effect wins.

$$\frac{\Delta\eta}{H} \simeq \left(\alpha_T \langle T - T_o\rangle - \beta_S \langle S - S_o\rangle\right) \tag{9.14}$$

where, as before, the angle brackets denote $\langle ()\rangle = \frac{1}{(H+\eta)} \int_{-H}^{\eta} () \ dz$ and Eq.(9.5) has been used. From Fig.9.5 we estimate that $\langle T - T_o\rangle \simeq 10\,^\circ\mathrm{C}$ over the top km of the warm water lens of the subtropical gyres and, from Fig.9.6, $\langle S - S_o\rangle \simeq 0.5\mathrm{psu}$. Thus if (see Table 9.4) $\alpha_T = 2\times10^{-4}\,^\circ\mathrm{C}^{-1}$ and $\beta_S = 7.6\times10^{-4}\mathrm{psu}^{-1}$ we find that:

$$\frac{\Delta\eta}{H} \simeq \left(\underbrace{2}_{temp} + \underbrace{-0.38}_{salt}\right) \times 10^{-3}$$

One may ask: do the subtropics have to be lighter? Let's consider the case where water in the subtropics is actually heavier. Then from the thermal wind relation, we will strong currents in the abyss. Because of the jagged topography, these currents will be damped and water at the ocean floor will move away from the high pressure. This brings down the isopycnals. For a stably stratified fluid, this makes the subtropics lighter.

The thermal wind relation is widely used to infer ocean currents from temperature and salinity measurements, because ocean currents, being relatively slow, are difficult to measure. For this purpose, one needs to assume a reference level where the currents are known. Traditional practice is to assume a level of no motion in the abyss. Now we have good satellite altimetry data, one could use the surface as that level.

Like the atmosphere, the ocean is full of eddies. These eddies are smaller compared to those in the atmosphere (why?). Variations due to these eddies are comparable to those of the mean:

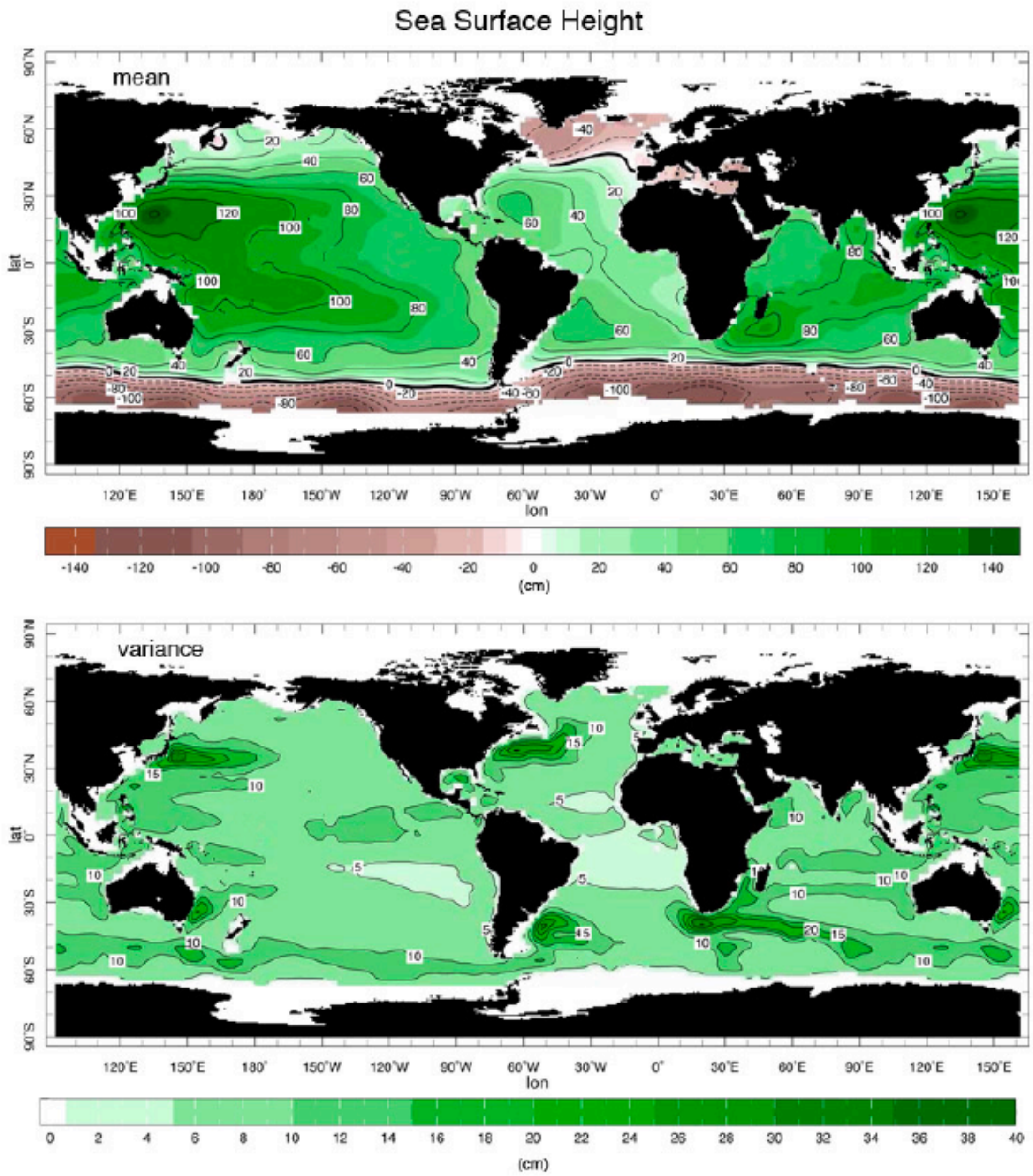


Figure 9.19: (Top) The 10-year mean height of the sea surface relative to the geoid, $\overline{\eta}$, (contoured every 20 cm) as measured by satellite altimeter. The pressure gradient force associated with the tilted free surface is balanced by Coriolis forces acting on the geostrophic flow of the ocean at the surface. Note that the equatorial current systems very evident in the drifter data — Figs. 9.14 and 9.22 — are only hinted at in the sea surface height. Near the equator, where f is small, geostrophic balance no longer holds. (bottom) The variance of the sea surface height, $\sigma_\eta = \sqrt{\overline{\eta'^2}}$, Eq.(9.17), contoured every 5 cm.

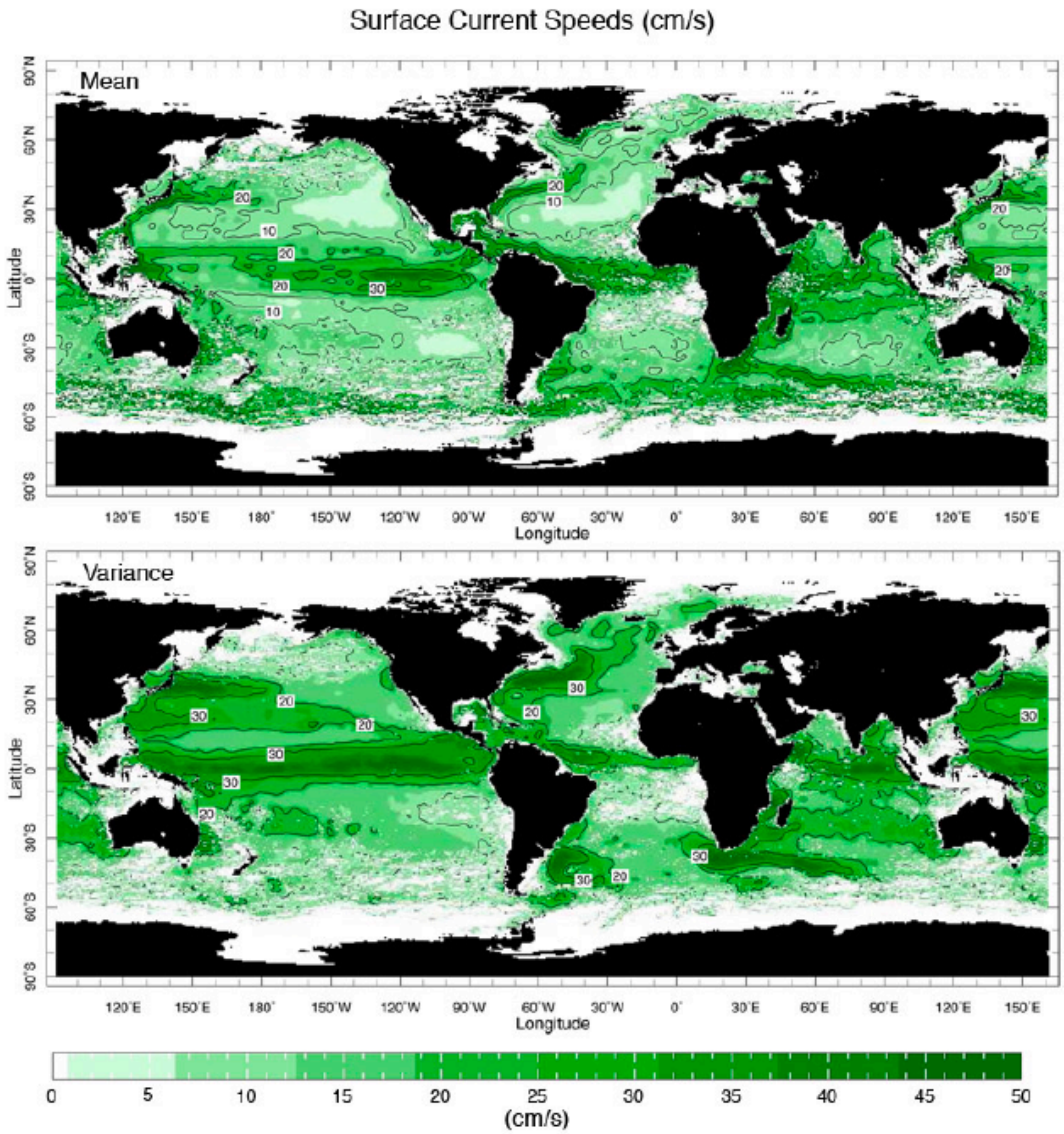


Figure 9.22: (Top) Mean surface drifter speeds, $\sigma_{\overline{u}} = \sqrt{(\overline{u}^2 + \overline{v}^2)^{\frac{1}{2}}}$, and (bottom) eddy drifter speeds, $\sigma_{u'} = \sqrt{\left(\overline{u'^2} + \overline{v'^2}\right)^{\frac{1}{2}}}$ computed from 20 y of drifter observations. Regions of the ocean in which observations are sparse — particularly in the southern oceans around Antarctica — appear as white gaps. Data courtesy of Maximenko and Niiler (personal communication, 2003).

So far we have tied the mean currents and slopes of the isopycnal surfaces to the free surface height. However, how do we determine the free surface height in the first place?