The Thermohaline Circulation

The wind-driven circulation that we described before is mechanically driven. Part of the oceanic circulation is also driven by water density variations caused by temperature and salinity variations. This is known as the thermohaline circulation. In nature, the two are strongly coupled and are not separable. But it is convenient (and has become traditional) to discuss them separately, and in general the wind-driven circulation dominates the currents near the surface while the density driven circulation dominates the flow in the abyss.

A number of processes modify the density of surface water: surface radiation and heat fluxes, evaporation and precipitation, formation and melting of ice. If these processes make water dense enough, water will sink in the form of oceanic convection. This is the same kind of thermal convection as in atmospheric convection, except here convection tends to occur at high latitudes where surface water is most dense and vertical stratification is most weak.

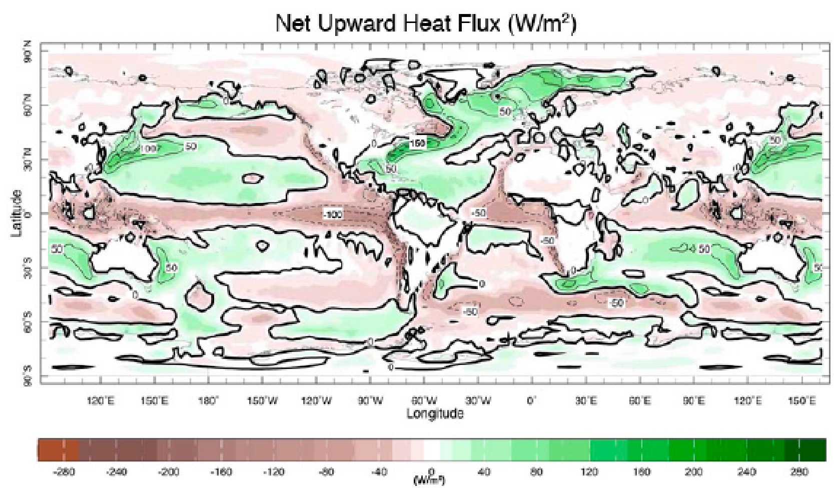
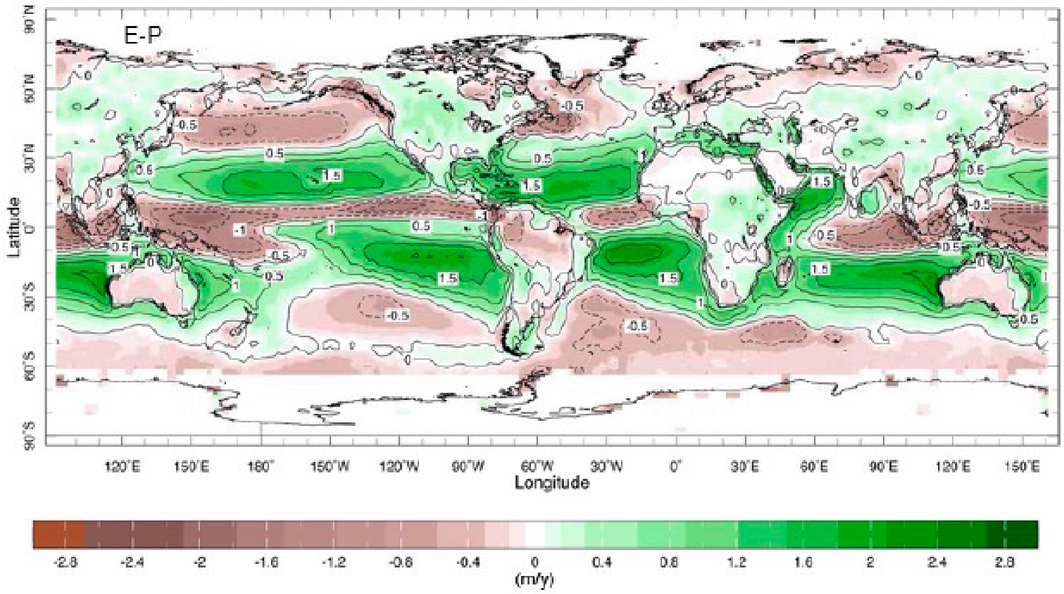
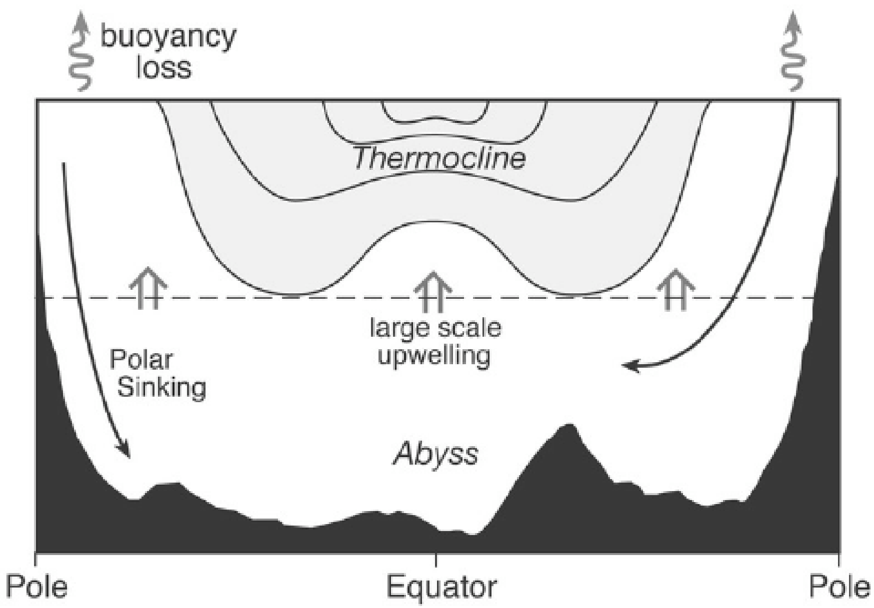


Figure 11.4: Global map of net annual-mean, constrained, heat flux, $\mathcal{Q}_{\text{net}}$, across the sea surface in Wm^{-2}. Areas in which the fluxes are upward, in to the atmosphere, are shaded green; areas in which the flux is downward, in to the ocean, are shaded brown. Contour interval is 50 Wm^{-2}. From Kalnay et.al. 1996.





Oceanic deep convection occurs only in limited regions in the ocean. Unlike in the atmosphere, where deep convection can be easily located because of the associated clouds, one need to measure the T,S profiles in the ocean to determine the depth of the mixed layer and whether deep convection is occurring. Regions with deep convection tend to have harsh environments, and it is a challenge to make measurements there.

Figure 11.10: The Woods Hole ship KNORR cuts through harsh Labrador Sea conditions during the winter Labrador Sea Deep Convection Experiment (Feb–Mar 1997) taking observations shown in Fig.11.11. Waves such as those shown at the top caused continual ice build-up on the ship, as can be seen at the bottom. Courtesy of Bob Pickart, WHOI.

Here is an example of measured ocean convection:

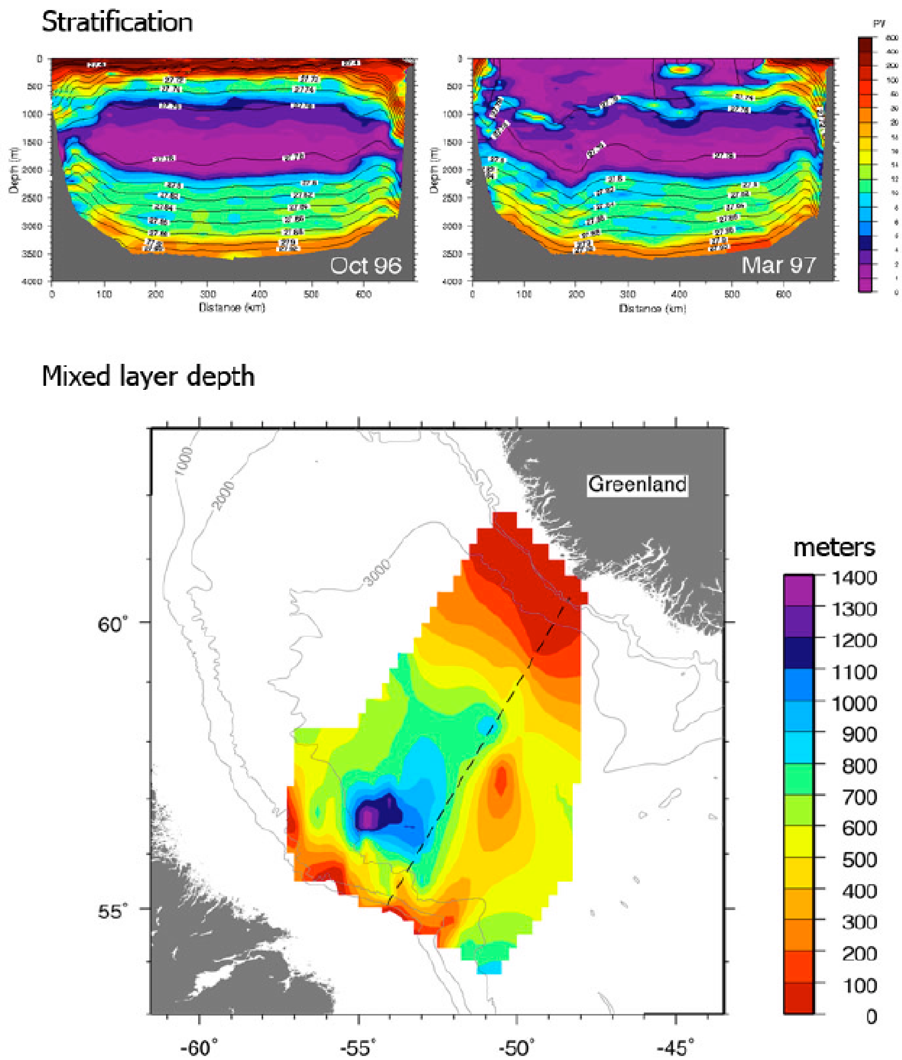


Figure 11.11: Top: Sections of potential density, σ (contoured) and stratification, $\frac{\partial\sigma}{\partial z}$ (colored) across the Labrador Sea in October 1996, prior to the onset of convection and in March 1997, after, and during wintertime convection. Purple indicates regions of very weak stratification. Bottom: A horizontal map of mixed layer depth observed in Feb-Mar 1997, showing convection reaching to depths in excess of 1 km. The position of the sections shown at the top is marked by the dotted line. Courtesy of Robert Pickart, WHOI.

And here is schematic of ocean convection. One of the differences between oceanic convection and atmospheric convection is that oceanic convection happens at high latitude thus has small deformation radius. So the spreading by gravity wave is not efficient.

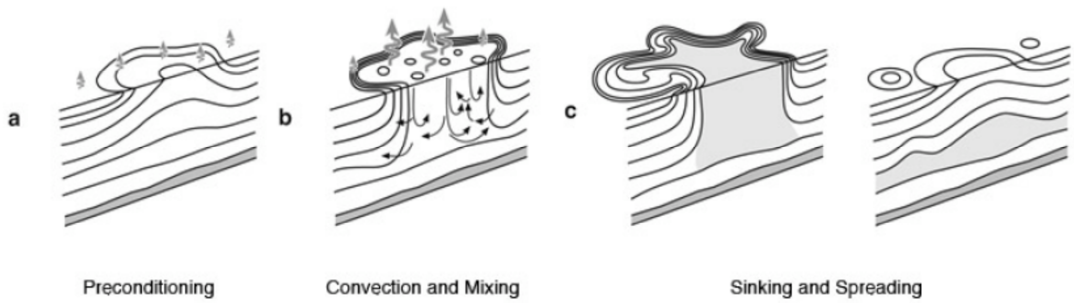


Figure 11.12: Schematic diagram of the three phases of open-ocean deep convection: (a) preconditioning, (b) deep convection and mixing and (c) sinking and spreading. Buoyancy flux through the sea surface is represented by curly arrows, and the underlying stratification/outcrops are shown by continuous lines. The volume of fluid mixed by convection is shaded. From Marshall and Schott (1999).

Such measurements by oceanographers suggest that convection reaches down into the abyssal ocean only in the North Atlantic (in the Labrador and Greenland seas) and in the South Atlantic (in the Weddell sea). These sites thus set and maintain the properties of the abyss. These regions are characterized by strong buoyancy loss, weak stratification, and cyclonic wind stress (thus Ekman suction). There is no deep convection in the North Pacific, which is considerably fresher.

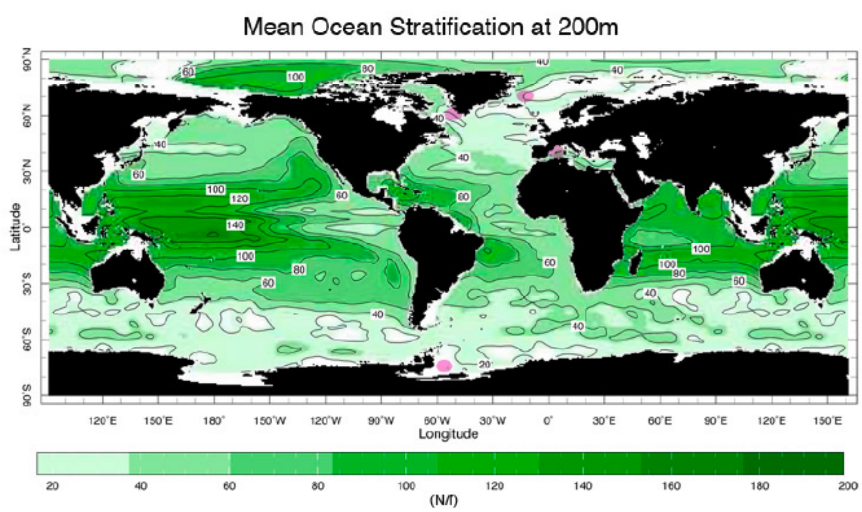
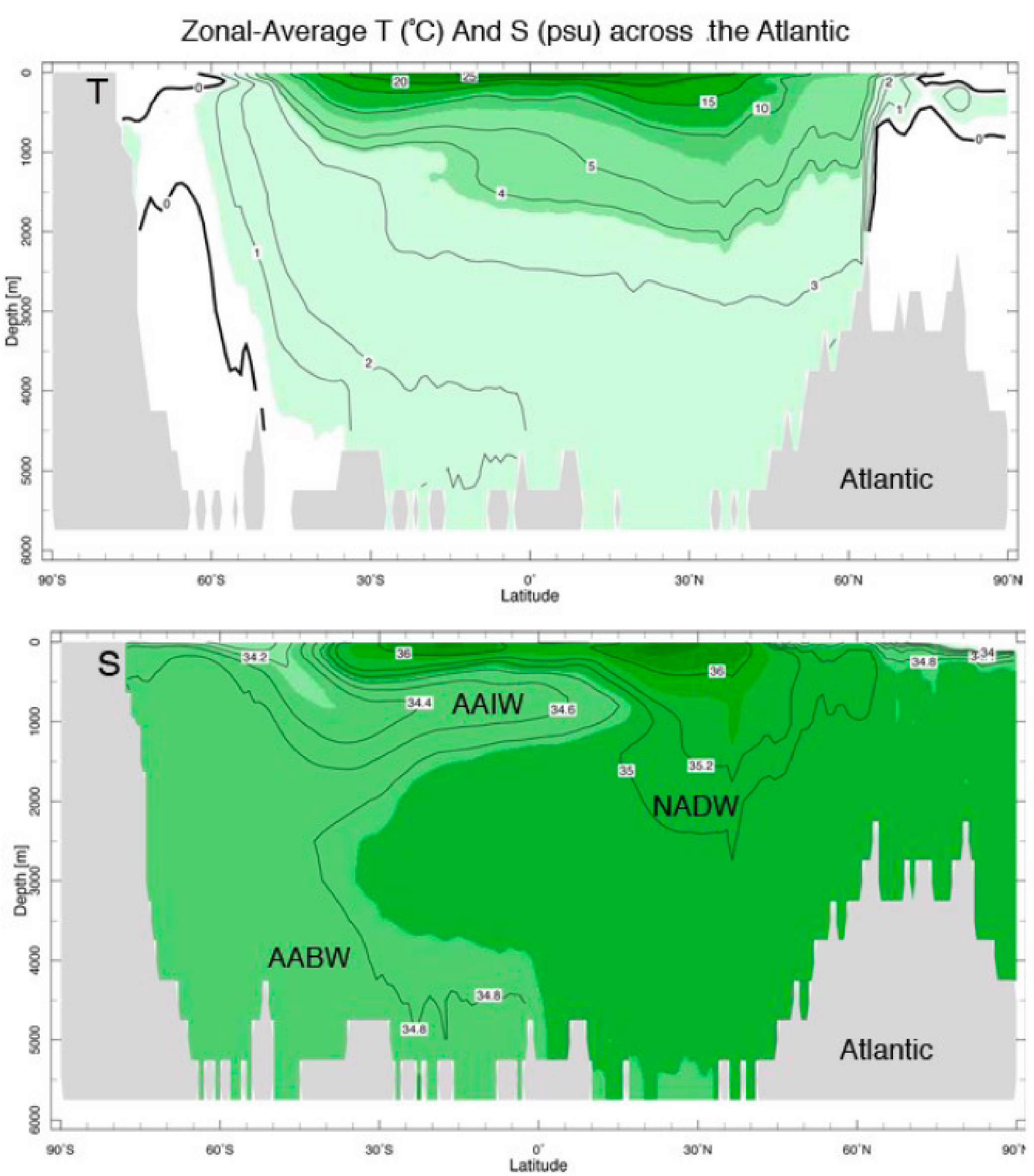


Figure 11.9: The annual mean stratification of the ocean at a depth of 200 m, as measured by $\frac{N}{f_{ref}}$: i.e. the buoyancy frequency, Eq.(9.6), normalized by a reference value of the Coriolis parameter, $f_{ref}=10^{-4}\,\mathrm{s}^{-1}$. Note that $\frac{N}{f_{ref}} \lesssim 20$ in regions where deep mixed layers are common — cf. Fig.9.10. Sites of deep-reaching convection are marked in the Labrador Sea, the Greenland Sea, the Western Mediterranean and the Weddell Sea.

Circulation in the abyss is weak and difficult to measure. Distributions of tracers have commonly been used to infer the abyss circulation. These include temperature, salinity, which only change slowly in the deep ocean, dissolved oxygen, CFCs, carbon-14, tritium (^{3}H) from atomic bomb tests, among others.



igure 11.13: Zonal average (0° $\longrightarrow$ 60° W) temperature (top) and salinity ›ottom) distributions across the Atlantic Ocean. Antarctic Intermediate Wa- r (AAIW), Antarctic Bottom Water (AABW) and North Atlantic Deep Water ʿADW) is marked. Compare this zonal-average section with the hydrographic ction along 25° W shown in Fig.9.9.

From the T, S distributions, one sees the different water masses. The deep water formation in the North Atlantic is believed to be caused by cooling of warm and saline water carried north by the Gulf Stream. Surface water around Antarctica however is fresher because of the excess of precipitation over evaporation. And the formation of Antarctic Bottom Water is more dependent on the brine rejection from sea ice formation. Dissolved oxygen is another useful tracer. Surface waters are near saturation in oxygen content (slightly supersaturated in fact, likely because of bubbles and photosynthesis in surface waters). As the water leaves the surface, its oxygen content is slowly used up by biological activity.

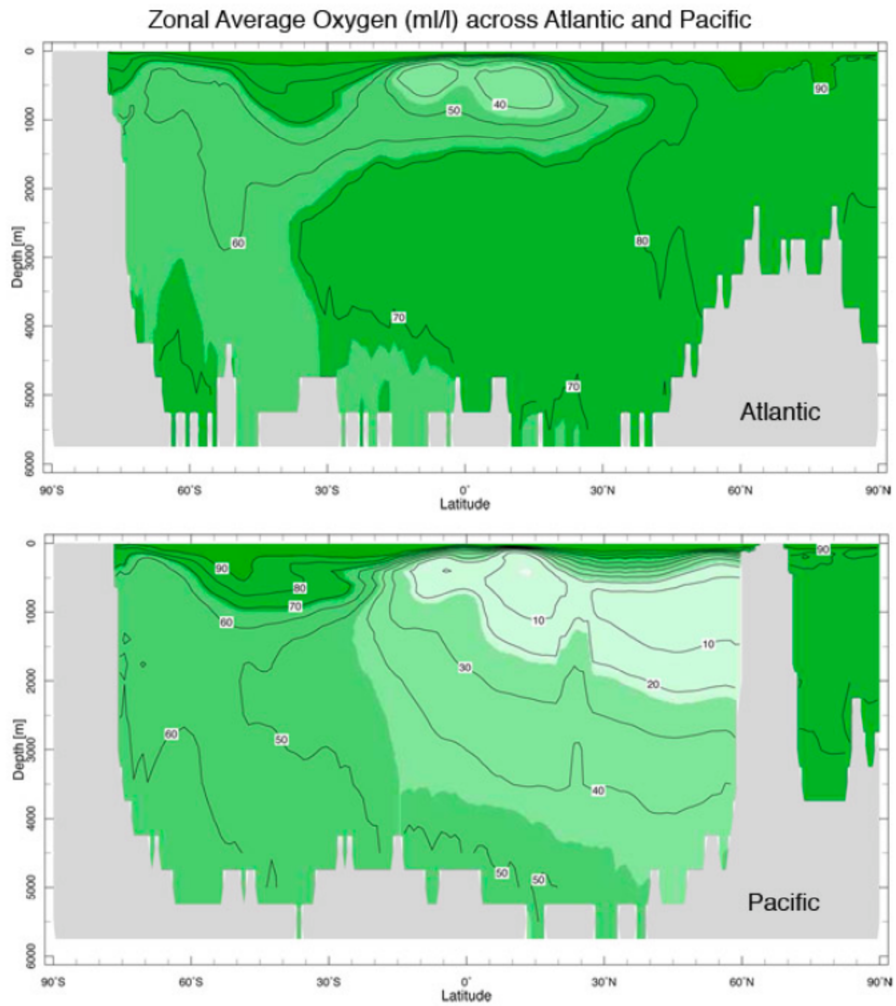


Figure 11.14: Zonally averaged oxygen saturation (in $\mathrm{ml\,l^{-1}}$) in the Atlantic ($0°-60°$W) and Pacific ($150-190°W$) oceans.

Measurements above (in milliliter per liter) give a sense of the direction of the circulation. Oxygen is not a great clock because how fast it decreases depends on biological activity and is quite uncertain. Carbon-14 (half time 5700 years) is more useful in this regard. Combining various measurements, it is estimated that the time it takes to replace the water in the deep ocean through convection in the regions of deep water formation is on the order of 1000 years. We may call this the turnover time of the ocean. The ocean therefore provides a long memory in terms of heat and chemical species.

It is interesting to contrast oceanic convection and atmospheric convection. Atmospheric convection is driven by heating at the surface and radiative cooling in the atmosphere. For ocean convection, surface cooling and increase in salinity provides the surface buoyancy fluxes but what is its version for radiative cooling? In the atmosphere, without radiative cooling, the atmosphere will eventually warm up and become stable to convection. By the same token, there needs to be some processes that warm the bottom ocean. Otherwise, it will also eventually become dense enough and stable to all convection. This was a big problem for oceanographers. While not completely resolved, it is believed that a combination of wave breaking and tides, particularly in regions of rough bottom topography can provide enough mixing that mixes heat downward to warm up the bottom ocean.

The ocean is important in many aspects. One of them is the transport of heat. The spatial scales of the motions important for transporting heat in the ocean are small (recall the small deformation radius in the ocean). It is not possible to simultaneously measure current and temperature with a sufficient sampling to directly infer the oceanic heat transport. Existing estimates with such an approach are very uncertain. Alternatively, the ocean heat transfer is inferred as the difference between total required heat transfer and the atmospheric heat transfer; the former can be estimated by considering top of the atmosphere radiative fluxes. One could also estimate ocean heat transport using surface radiative and heat fluxes.

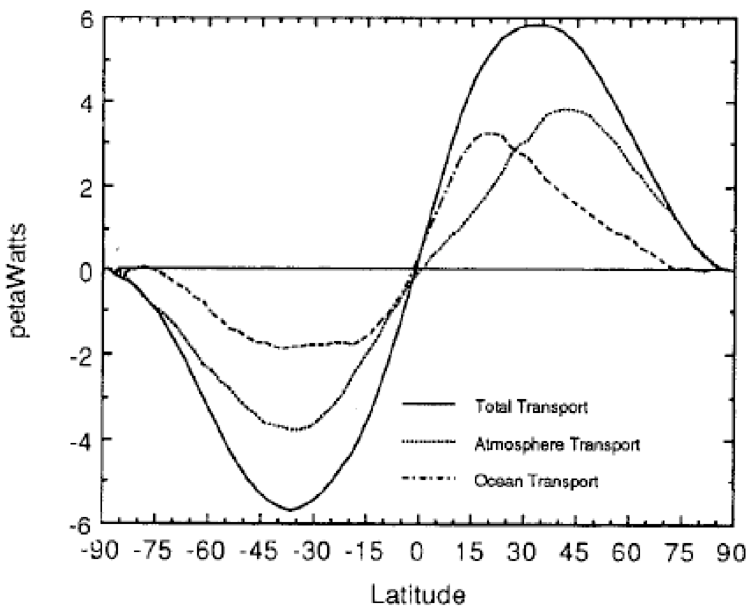


Fig. 7.16 Estimates of the annual mean meridional energy transport required by the energy balance at the top of the atmosphere, estimated from observations in the atmosphere, and the oceanic transport obtained by subtracting the atmospheric energy transport from the total transport required by the annual energy balance (7.16). Net transport inferred from Earth Radiation Budget Experiment data. [Atmospheric transport data from Peixóto and Oort (1984). Used with permission from the American Physical Society.]

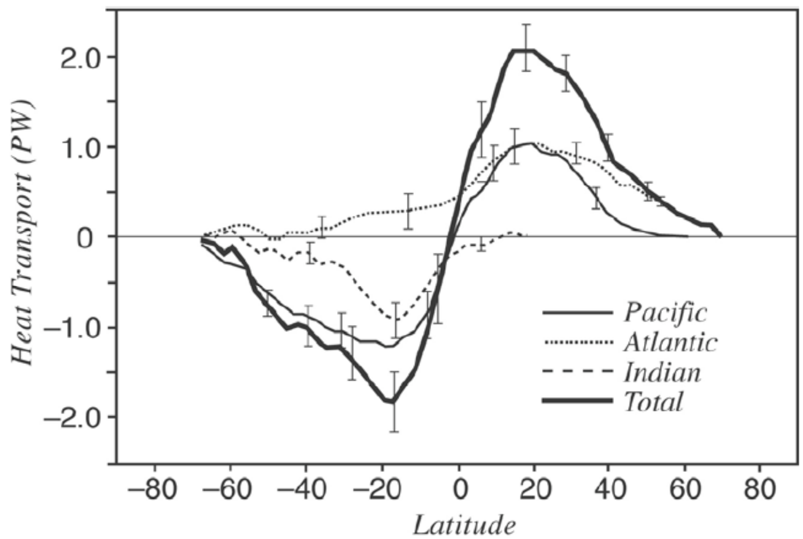


Figure 11.27: Northward heat transport in the world ocean, $\overline{\mathcal{H}}^{\lambda}_{ocean}$, and by ocean basin calculated by the residual method using atmospheric heat transport from ECMWF and top of the atmosphere heat fluxes from the Earth Radiation Budget Experiment satellite. The vertical bars are estimates of uncertainty. From Houghton et al. (1996) using data from Trenberth and Solomon (1994).

Both the wind-driven circulation and the thermohaline circulation are believed to be important in transporting heat. Ocean eddies are very small in size. It's not clear if they are important to the ocean transport except in the Antarctic circumpolar current.