Atlantic climate: The NAO and the AMOC

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2.1 Atmospheric mean state

To understand what the NAO is, we first need to understand the atmospheric mean state over the Atlantic and north Pacific during the summer (JJA) and the winter (DJF). Figure 1 shows the mean sea level pressure during winter and summer. In the winter, when the ocean is considerably warmer than the continents, the sea level pressure is high over Asia and low over the north Pacific and north Atlantic. Note the weak high pressure center just south of the Icelandic Low in DJF. During JJA this high pressure center intensifies and becomes the main feature in the north Atlantic (see Fig. 1).

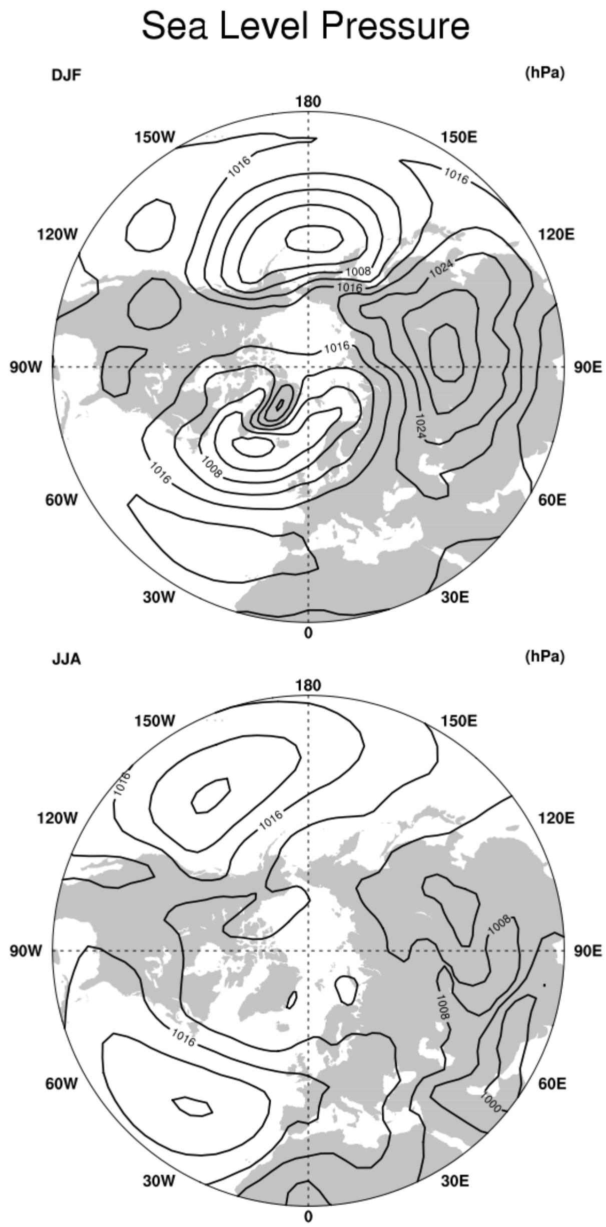

Figure 1. Mean sea level pressure for (top) boreal winter (December-February) and (bottom) boreal summer (June-August). The data come from the NCEP/NCAR reanalysis project over 1958-2001 [*Kalnay et al.,* 1996], and the contour increment is 4 hPa.

Figure 1: Figure credit: (Hurrell et al., 2003)

Remember that in the northern hemisphere flow moves clockwise around high-pressures and anti-clockwise around low pressures (see Fig. 2), which demonstrates why the flow at the surface over the north Atlantic is from west to east, running from the northeast US to Europe. Figure 2 also shows how the northern hemisphere jet stream at 200 hPa is stronger in the winter than the summer.

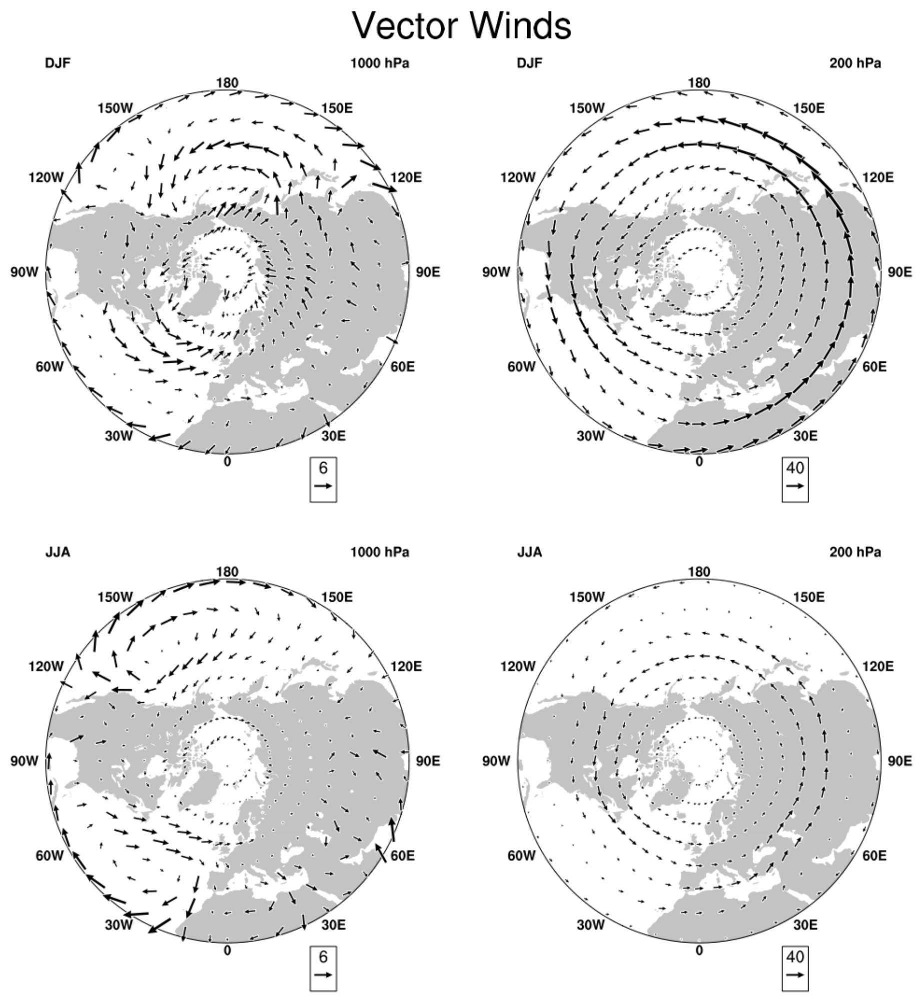

Figure 2. Mean vector winds for (top) boreal winter (December-February) and (bottom) boreal summer (June-August) for (left) 1000 hPa and (right) 200 hPa over 1958-2001. The scaling vectors are indicated in the boxes and are given in units of m s^{-1}.

Figure 2: Figure credit: (Hurrell et al., 2003)

2.2 North Atlantic Oscillation (NAO)

Whether the NAO is a distinct mode of variability or if it is instead the north Atlantic expression of the hemispheric Pacific North American (PNA) pattern is still debated. Figure 3 illustrates both patterns

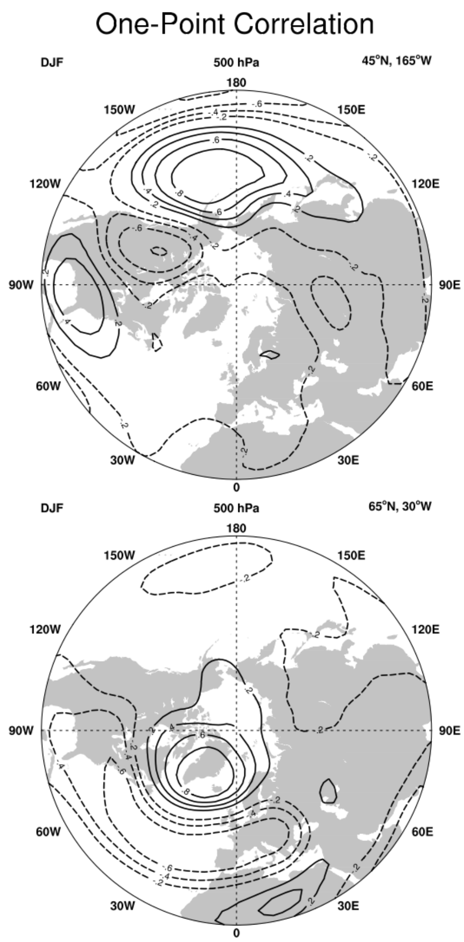

Figure 5. One-point correlation maps of 500 hPa geopotential heights for boreal winter (December-February) over 1958-2001. In the top panel, the reference point is 45ºN, 165ºW, corresponding to the primary center of action of the PNA pattern. In the lower panel, the NAO pattern is illustrated based on a reference point of 65ºN, 30ºW. Negative correlation coefficients are dashed, the contour increment is 0.2, and the zero contour has been excluded.

Figure 3: Figure credit: (Hurrell et al., 2003)

The PNA has four centers of action expressed as geopotential height anomalies, which pivot along the mean position of the Pacific jet stream. Physically, the PNA pattern represents changes in the north-south migration of the atmospheric flow (also referred to as its 'waviness'). These changes in the position of the jet stream are linked to ENSO on interannual timescales, and therefore reflect the response of the higher latitudes to deep convection in the tropics. Past studies have demonstrated, however, that ENSO isn't the only phenomena capable of producing the PNA, which implies that it's a mode intrinsic to the atmosphere.

Similar to the PNA, the NAO represents the position and strength of the atmospheric flow over the Atlantic. And regardless of whether it's separate from the PNA or not, we can describe how its high and low phases affect the climate in the northern hemisphere. When the NAO is in its 'positive' phase, the storm track is more intense and less 'wavy', making northern Europe and the southern US warm and wet.

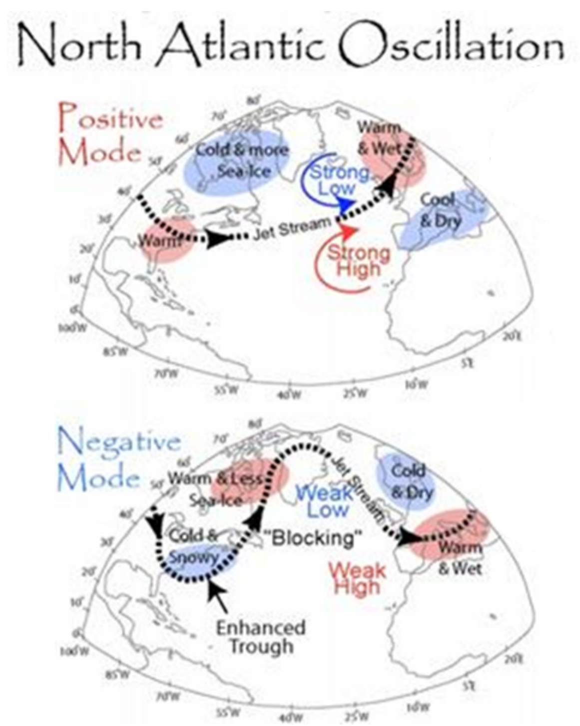

Figure 4: Cartoon of the positive and negative phases of the NAO.

Source:http://airmap.unh.edu/background/nao.html#nao

2.3 Temporal characteristics

Unlike ENSO, the NAO often changes sign from one month to the next. As a mode of variability inherent to the atmosphere (rather than being forced by ocean SST anomalies) it operates on the time-scales of the atmosphere (see Fig. 5. Similar to ENSO, however, the NAO is strongest in boreal winter, when baroclinic instability is strongest in the northern hemisphere. One final note about the NAO: despite it's monthly variability, and despite being a mode inherent to the atmosphere, the NAO exhibits variability on decadal timescales, which implies that either the ocean or the stratosphere may modulate the NAO, although this is an ongoing area of research.

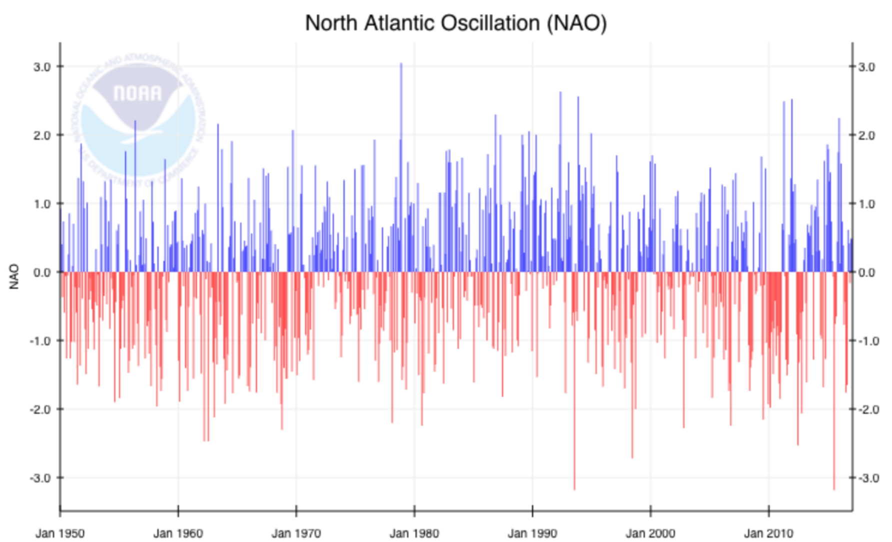

Figure 5: Monthly index of the NAO.

Source:`https://www.ncdc.noaa.gov/teleconnections/nao/`

3 The Atlantic ocean

3.1 Atlantic Meridional Overturning Circulation (AMOC)

The Atlantic basin is unique in that the meridional circulation of the ocean can be described as a single coherent overturning cell, the Atlantic meridional overturning circulation (AMOC). Central to this overturning cell is the location of oceanic convection (deep water formation) and the location of upwelling. Deep water formation occurs in the North Atlantic, where salty water from the Atlantic mix with cold water from the pole to form dense water that sinks and turns equatorward. Upwelling primarily occurs in the Southern Ocean via flow along isopycnals as they outcrop to the surface due to the strong winds around

Antarctica. Figure 6 depicts flow in (a) the surface layer and (b) the deep layer of the Atlantic that results from a combination of the wind-driven gyres in the mid-latitudes, and the internal mode of deep water formation and upwelling. Figure 7 is a more realistic depiction of this same circulation, although the basics are the same.

These two primary modes reflect the observed timescales of variability. On interannual timescales variability in meridional transport (at least as far as we've observed) is connected to variability in the wind stress forcing. On decadal to centennial timescales, the variability in AMOC is a complex combination of wind-driven processes and density-drive processes. The mean winds and the NAO are of obvious importance for the wind-driven circulation. The density-driven deep water formation occurs in only a few specific locations of the north Atlantic, such as the Labrador sea, where the subtropical and subpolar gyres meet. To see how deep-water formation affects meridional flow, consider the Taylor-Proudman effect

$$\beta v = f\frac{w_{ek}}{h}$$

where $w_{ek} < 0$ is Ekman pumping (i.e. deep water formation) that, by conservation of potential vorticity, leads to equatorward flow ($v < 0$). This explains why deep-water formation in the North Atlantic leads to deep western boundary currents

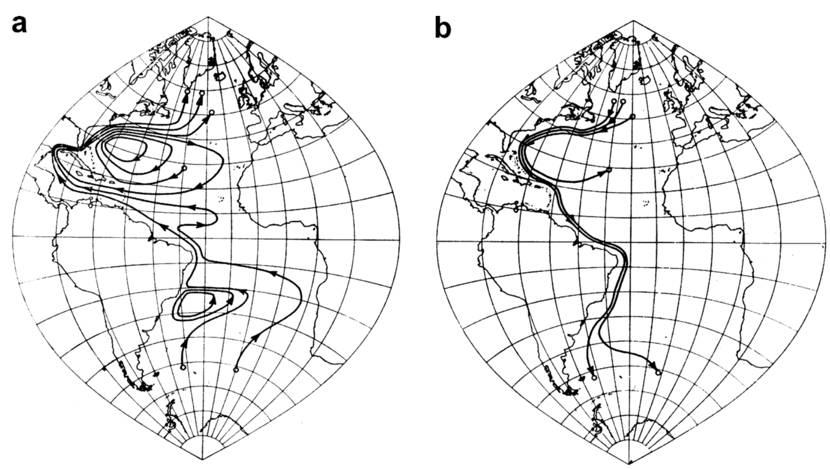

Fig. 11. Schematic upper layer (a) and deep layer (b) circulation in the Atlantic including a superposition of the internal mode and wind-driven components as shown by Stommel (1957). Small circles indicate general regions of vertical motion connecting surface and deep layers.

Figure 6: Superposition of the wind-driven circulation and the internal mode in the Atlantic

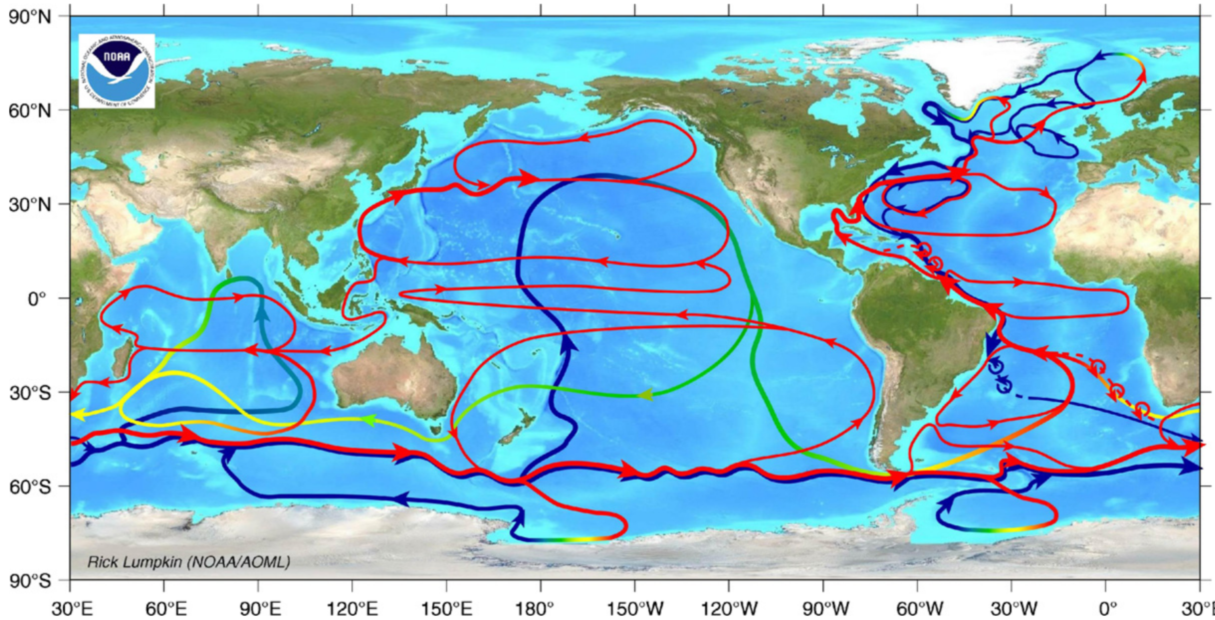

Fig. 19. Schematic diagram of the pathways of the global conveyor belt circulation and some of its recirculating loops provided by Lumpkin (personal communication). Red arrows indicate light, upper (surface) water, blue arrows are bottom and deep water. Two new features are included: (1) significant westward return flow south of Australia, and (2) large ocean eddies in the Atlantic transporting part of the meridional overturning circulation.

Figure 7: A more complete view of the global ocean 'conveyor belt'

Another way to visualize the 3-D flow in the Atlantic is pictured in Figure 8, which illustrates how the surface flow relates to the meridional flow as a function of height. Deep water is formed in the North Atlantic and Antarctica. Deep water in Antarctica is generally described as either Antarctic intermediate water (AAIW) or Antarctic bottom water (AABW) depending on the depth to which it sinks. Deep water formed in the North Atlantic is referred to as north Atlantic deep water (NADW).

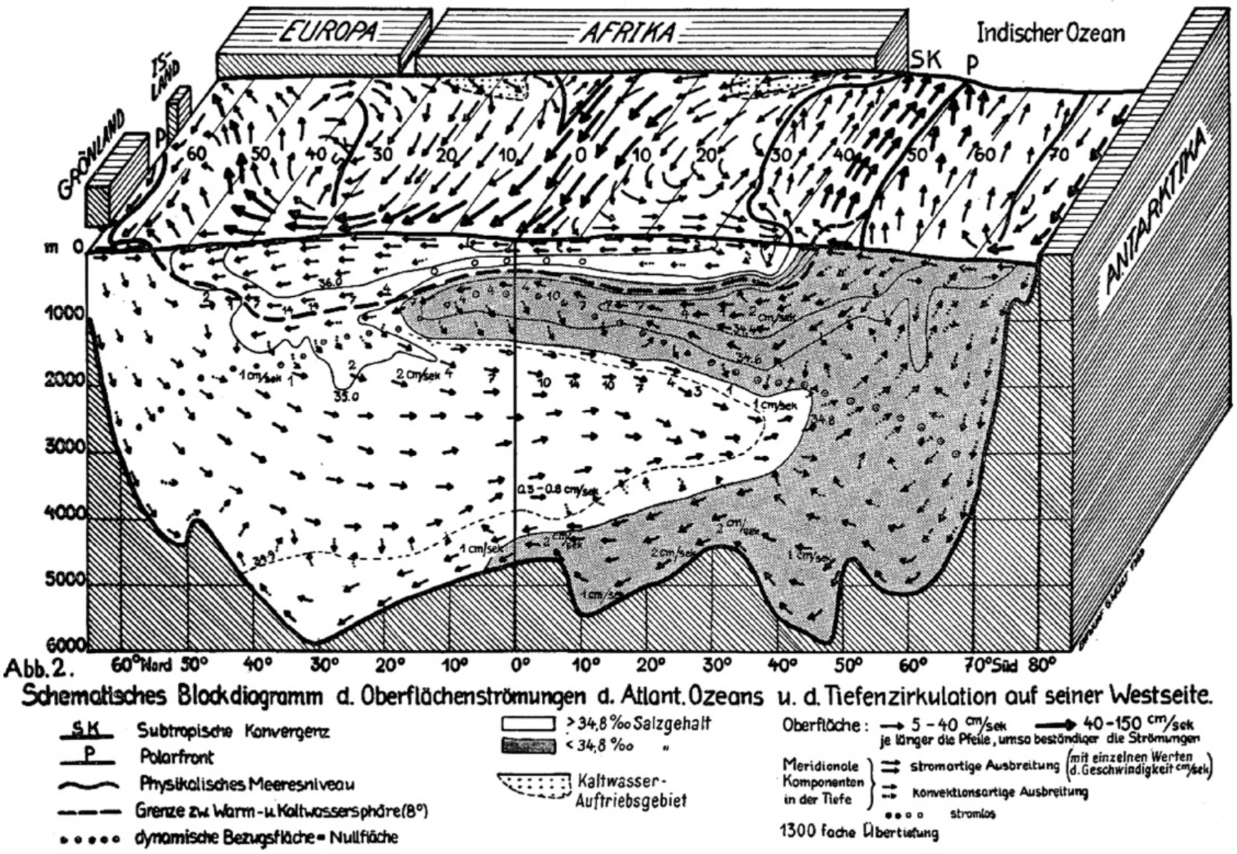

Fig. 7. Block diagram of surface currents, salinity contours and the meridional circulation along a section in the western Atlantic by Wüst (1949) based on the Meteor expedition and other hydrographic data (Wüst, 1935). Small numbers are geostrophic speeds in cm/s. Dashed line is the boundary (9 °C) between warm and cold layers.

Figure 8: A vertical slice of the Atlantic meridional circulation

3.2 The thermohaline circulation

To understand how the AMOC transports heat, we need to look at ocean-atmosphere fluxes. Figure 9 shows where the ocean gains heat, and where it loses heat. The ocean loses heat to the atmosphere in the western boundary currents (particularly the Gulf Stream and the Kuroshio Current). The ocean gains heat along the equatorial cold tongue and in regions of upwelling. In the Atlantic at the surface, water moves northward from the southern hemisphere up to the north Atlantic, gaining heat all the while. Once in the north Atlantic the water rapidly loses heat to the atmosphere, sinks, and moves equatorward. Through this process the Atlantic is a major source of cross-equatorial heat transport (see Figure 10). We refer to the transport of heat and salt in the Atlantic as the thermohaline circulation. It's hypothesized that the thermohaline circulation is the primary reason why the northern hemisphere atmosphere is warmer than the southern hemisphere, and therefore why the ITCZ is located (on average) north of the equator.

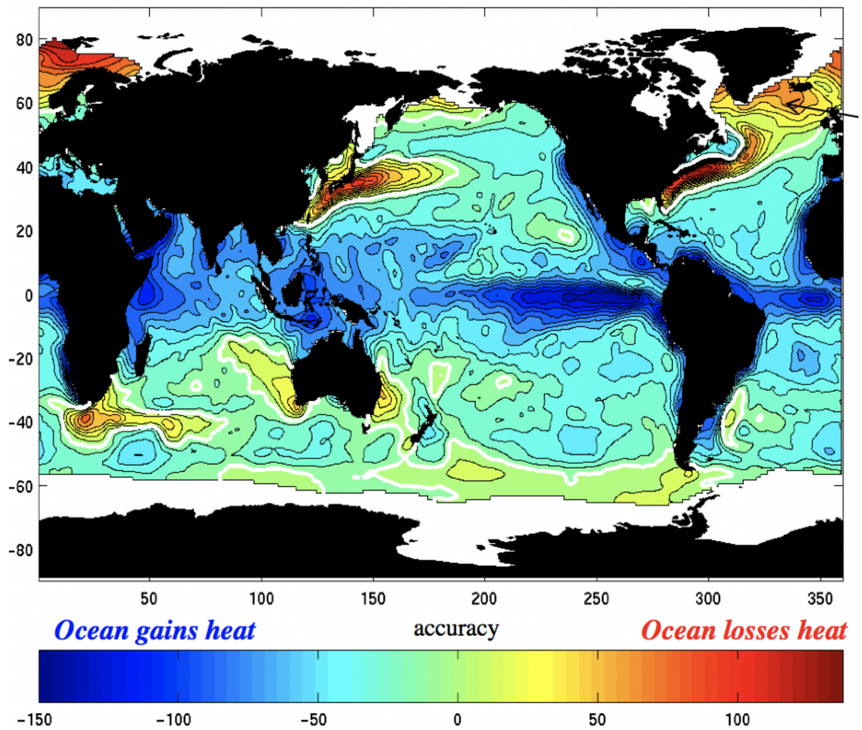

Figure 9: Fluxes of heat into and out of the ocean. Note that in the Atlantic the ocean gains heat over the south and central Atlantic, then loses that heat in the Gulf Stream and in the North Atlantic, where water convects (sinks, mixes and begins traveling equatorward)

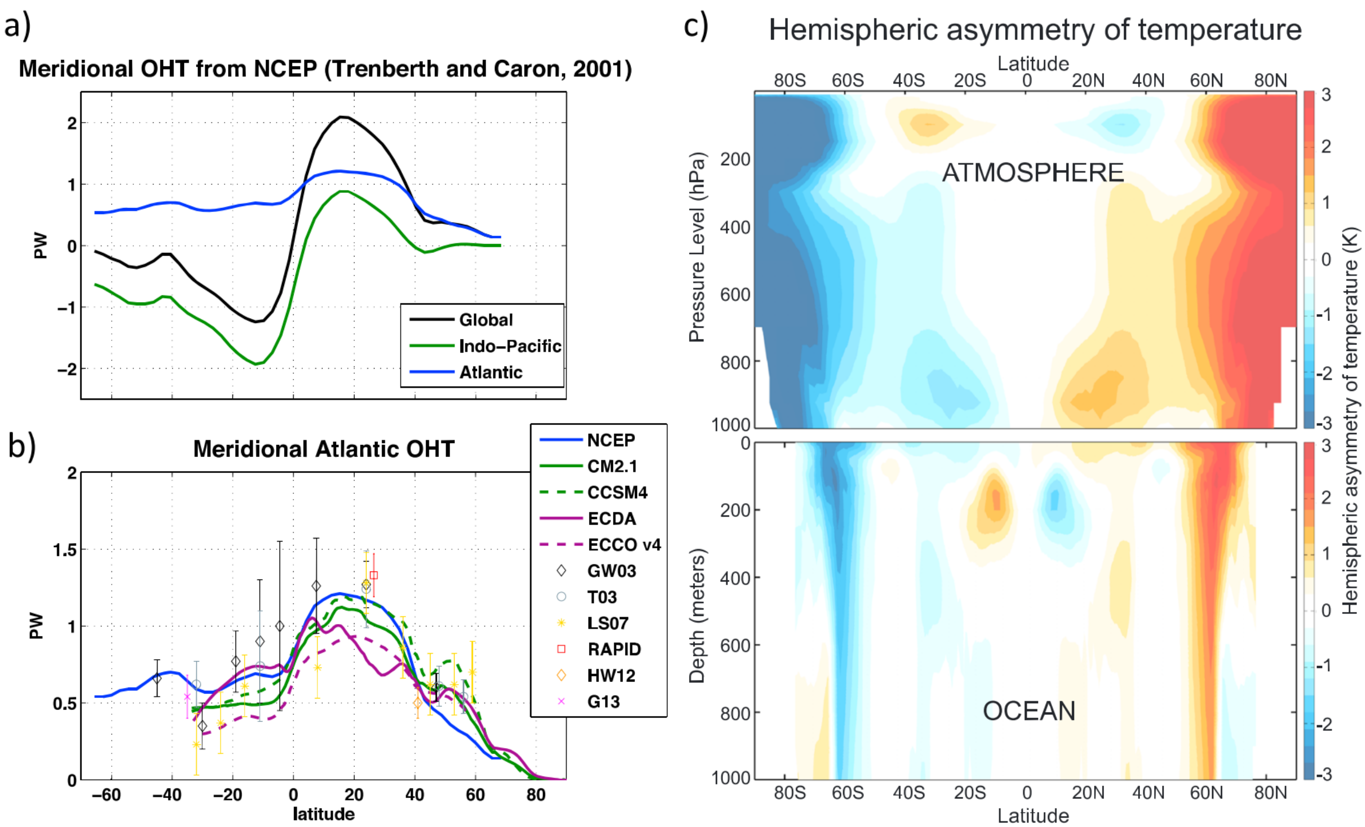

Figure 3. (a) Meridional ocean heat transports (OHT, positive northward) in PW (10^{15} W) for the global ocean (black), the Indo-Pacific (green), and the Atlantic (blue) from NCEP atmospheric reanalysis [*Trenberth and Caron*, 2001]. (b) Atlantic OHT from NCEP atmospheric reanalysis (blue) [*Trenberth and Caron*, 2001] is compared to several direct estimates: *Ganachaud and Wunsch* [2003] (black diamonds), *Talley* [2003] (grey circles), *Lumpkin and Speer* [2007] (yellow stars), the RAPID-MOCHA array at 26.5°N (red square) [*Johns et al.*, 2011], *Hobbs and Willis* [2012] (orange diamond), and *Garzoli et al.* [2013] (magenta cross). The vertical bars indicate the uncertainty range for the direct estimates. Also compared are the Atlantic OHT in CM2.1 (green solid) and CCSM4 (green dashed) preindustrial control simulations (modified from *Msadek et al.* [2013]) and the GFDL ECDA (1960–2010, solid purple) [*Chang et al.*, 2012] and ECCO v4 (1992–2012, dashed purple) [*Forget and Ponte*, 2015; *Forget et al.*, 2015] ocean state estimates. (c) Observed hemispheric asymmetry of temperature in the atmosphere and ocean (in °C) computed from the NCEP reanalysis and the World Ocean Atlas. The asymmetric component of a temperature field $T(\phi)$ is defined as $T_{as}(\phi)=(T(\phi)-T(-\phi))/2$ where ϕ is the latitude [from *Marshall et al.*, 2014a].

Figure 10: Meridional ocean heat transport in each ocean basin. Panel c shows the asymmetries between the hemispheres, in particular this panel highlights how northward heat transport by AMOC may contribute to the NH atmosphere being warmer than the SH

References

James W Hurrell, Yochanan Kushnir, Geir Ottersen, and Martin Visbeck. *An overview of the North Atlantic oscillation.* Wiley Online Library, 2003.